Ice streams in the Laurentide Ice Sheet: identification, characteristics and comparison to modern ice sheets

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Abstract:

This paper presents a comprehensive review and synthesis of ice streams in the Laurentide Ice Sheet (LIS) based on a new mapping inventory that includes previously hypothesised ice streams and includes a concerted effort to search for others from across the entire ice sheet bed. The inventory includes 117 ice streams, which have been identified based on a variety of evidence including their bedform imprint, large-scale geomorphology/topography, till properties, and ice rafted debris in ocean sediment records. Despite uncertainty in identifying ice streams in hard bedrock areas, it is unlikely that any major ice streams have been missed. During the Last Glacial Maximum, Laurentide ice streams formed a drainage pattern that bears close resemblance to the present day velocity patterns in modern ice sheets. Large ice streams had extensive onset zones and were fed by multiple tributaries and, where ice drained through regions of high relief, the spacing of ice streams shows a degree of spatial self-organisation which has hitherto not been recognised. Topography exerted a primary control on the location of ice streams, but there were large areas along the western and southern margin of the ice sheet where the bed was composed of weaker sedimentary bedrock, and where networks of ice streams switched direction repeatedly and probably over short time scales. As the ice sheet retreated onto its low relief interior, several ice streams show no correspondence with topography or underlying geology, perhaps facilitated by localised build-up of pressurised subglacial meltwater. They differed from most other ice stream tracks in having much lower length-to-width ratios and have no modern analogues. There have been very few attempts to date the initiation and cessation of ice streams, but it is clear that ice streams switched on and off during deglaciation, rather than maintaining the same trajectory as the ice margin retreated. We provide a first order estimate of changes in ice stream activity during deglaciation and show that around 30% of the margin was drained by ice streams at the LGM (similar to that for present day Antarctic ice sheets), but this decreases to 15% and 12% at 12 cal ka BP and 10 cal ka BP, respectively. The extent to which these changes in the ice stream drainage network represent a simple and predictable readjustment to a changing mass balance driven by climate, or internal ice dynamical feedbacks unrelated to climate (or both) is largely unknown and represents a key area for future work to address.

1. Introduction

Ice sheets lose mass through melting or dynamically through discharge via rapidly-flowing ice streams/outlet glaciers. Recent studies of ice sheet velocity patterns have revealed an intricate network in Antarctica (Joughin et al., 1999; Rignot et al., 2011b) and Greenland (Joughin et al., 2010b), with major ice stream trunks fed by smaller tributaries that extend far into the ice sheet interior (Fig. 1). These ice streams account for approximately 90% of mass loss in Antarctica (Bamber et al., 2000) and approximately 50% in Greenland (van den Broeke et al., 2009). They typically exhibit flow velocities of the order of hundreds m a^{-1}, increasing towards several km a^{-1} towards some of their termini (Joughin et al., 2010b; Rignot et al., 2011b). The rapid velocity and low surface gradient that characterise some ice streams result from a weak bed of saturated, fine-grained sediments that cannot support high basal shear stresses and either deforms or permits basal sliding across the bed (Alley et al., 1986; Bentley, 1987; Bennett, 2003). In contrast, a large number of ice streams arise from thermo-mechanical feedbacks that generate increased ice velocity through large topographic troughs and which may also be the focus of sediments and basal meltwater (Payne, 1999; Truffer and Echelmeyer, 2003). These two types of ice streams have been referred to as 'pure' and 'topographic' (cf. Stokes and Clark, 2001; Bennett, 2003; Truffer and Echelmeyer, 2003), although, in reality, they represent end members of a continuum.

Fig. 1 here, *full page width*

Ice streams observed in modern-ice sheets show considerable spatial and temporal variability, with changes in their velocity observed over timescales of hours to decades (Bindschadler et al., 2003; Joughin et al., 2003); and with some ice streams known to have switched on and off, and others changing their flow trajectory (Retzlaff and Bentley, 1993; Conway et al., 2002). Such variability may arise from external forcing (e.g., changes in atmospheric or oceanic conditions) or internal forcing (e.g., the availability of lubricating water and till; see review in Bennett, 2003). Elucidating these controls is a key area of research due to the contribution of ice streams to sea level rise (Nick et al., 2013), but satellite measurements and geophysical surveying of modern day ice streams only span a period of several decades. As such, they only provide a snapshot view of the system and are unable to observe their longer-term behaviour, such as those related to major changes in ice sheet configuration and volume over centennial to millennial time-scales. However, palaeo-ice streams can be reconstructed from the landform and sedimentary record on former ice sheet beds (Stokes and Clark, 2001). Unimpeded access to former ice stream beds also facilitates investigation of their bed properties and enables a better understanding of the mechanics of ice stream motion and the processes that facilitate and hinder fast ice flow (Beget, 1986; Hicock et al., 1989; Stokes et al., 2007). Ice streams may also transport sediments over large distances and knowledge of mineral dispersal patterns is economically important for the mining industry (e.g. Klassen, 1997).

A large number of palaeo-ice streams have been described for the Laurentide Ice Sheet (LIS; Fig. 2), the largest of the ephemeral Northern Hemisphere ice sheets, covering the territory of present day Canada from the Cordillera to the Arctic and Atlantic oceans, with large lobes extending to the north-eastern part of the present day United States (Denton and Hughes, 1981; Winsborrow et al., 2004). Ice streams draining the LIS into the North Atlantic have also been identified as a source of ice rafted debris (IRD) found in the ocean sedimentary record (Bond et al., 1992). These layers of IRD on the ocean floor have been interpreted to document periods of significant dynamic mass loss from the Pleistocene ice sheets of the Northern Hemisphere (Heinrich events; Heinrich, 1988; Andrews, 1998), particularly, but not exclusively, in the vicinity of the Hudson Strait Ice Stream (MacAyeal, 1993; Andrews and MacLean, 2003; Hemming, 2004; Alley et al., 2005).

The large number of hypothesised ice streams in the LIS (Winsborrow et al., 2004), coupled with the evidence of major purges of the ice sheet (Heinrich events), highlights the potential impact of ice streams on large-scale ice sheet dynamics, but there remain key areas of uncertainity that limit our understanding and predictions of modern ice sheet dynamics. For example, our knowledge of the scale and magnitude of episodes of ice sheet collapses is still in its infancy (MacAyeal, 1993; Deschamps et al., 2012; Kleman and Applegate, 2014), and it is unclear whether ice streams might accelerate ice sheet deglaciation beyond that which might be expected from climate forcing alone. Tackling these issues requires a comprehensive understanding of the location and timing of ice streams in palaeo-ice sheets. Numerical modellling of ice streams also requires testing against palaeo-data (e.g., Stokes and Tarasov, 2010) to further increase our confidence in their ability to simulate future ice sheet dynamics.

With these issues in mind, this paper presents a comprehensive review and analysis of ice streams in the LIS. It builds on a recent mapping inventory of their location (Margold et al., in press; Fig. 2) and here we: (i) briefly describe the historical emergence of the phenomena known as 'ice streams' in relation to the LIS; (ii) review the evidence of ice streams from different sectors of the LIS; (iii) analyse their characteristics in terms of their size, shape, and setting; (iv) examine the controls on their spatial and temporal activity; and (v) discuss their wider role in LIS dynamics and stability. We also make comparisons with ice stream activity in modern ice sheets, particularly those in Antarctica, where ice sheet extent is similar to that of the LIS during the Last Glacial Maximum (LGM; cf. Figs. 1, 2). A detailed comparison of ice streaming in the LIS with a modern-ice sheet has not yet been made, and it is useful to examine whether the configuration of ice streams at different stages during deglaciation differs from the drainage patterns seen in a modern ice sheet.

Fig. 2 here, *full page width*

2. Historical Perspective on Ice Streams in the LIS

In relation to the LIS, ice streams were first mentioned in 1895 when Robert Bell inferred, on the basis of striae mapping, the existence of a "great ice stream" passing through Hudson Strait to the Atlantic (Bell, 1895, p. 352-353; Brookes, 2007). The term did not appear again in connection with the LIS until Løken and Hodgson (1971) concluded that ice streams were responsible for eroding deep troughs on the continental shelf off the northeast coast of Baffin Island (Fig. 3). This, and other occurrences of the term in relation to the LIS (e.g., Hughes et al., 1977; Sugden, 1977), coincided with the early work on Antarctic ice streams (e.g., Hughes, 1977) that began to describe the phenomenon of ice streaming and provided a basis in glacier physics.

As more knowledge was gained about Antarctic ice streams, the concept of the Pleistocene Northern Hemisphere ice sheets as dynamic complexes of ice domes and saddles emerged, and both ice streams and ice shelves were depicted and described in Denton and Hughes (1981). Soon after, and in relation to the reconstructions by Denton and Hughes (1981), Andrews (1982, p. 25) commented that "*it is not known whether or where ice streams existed in the Laurentide Ice Sheet*". However, the concept of ice streaming was clearly gaining traction, and Dyke and Prest (1987a, 1987c) included their location (marked as convergent flow-lines) in their seminal publications describing the Late Wisconsinan and Holocene history of the LIS. Nonetheless, scepticism remained, with Mathews (1991, p. 265) suggesting that "*with so little known about the conditions and processes operating at the bed of contemporary ice streams, it seems doubtful that the site of an ancient ice stream can be identified solely from a track engraved on the substratum*". Such pessimism was misplaced, because Dyke and colleagues had already identified evidence of several ice stream tracks on the islands and peninsulas of the central Canadian Arctic (Dyke et al., 1982; Dyke, 1984; Dyke and Morris, 1988), largely on the basis of carbonate rich tills dispersed through areas of igneous or metamorphic bedrock. These dispersal trains were clearly traceable not only in the field, but also in aerial photographs, due to the colour contrast of carbonate rich tills against darker coloured autochthonous bedrock (Fig. 4). Some of the large channels of the Canadian Arctic were also suggested to have hosted topographically constrained ice (Dyke and Prest, 1987a) and/or ice shelves (Dyke and Prest, 1987c) with many later confirmed by landform assemblages on islands adjacent to the major straits and sounds, e.g., Victoria Island, bordering Amundsen Gulf (Sharpe, 1988; Fig. 3).

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During the 1970s and 80s, the glacial geological record of the southern margin of the ice sheet was being heavily scrutinised. This coincided with the 'paradigm shift' in glaciology that recognised the importance of fine-grained, deformable sediments in facilitating fast ice flow (Boulton, 1986), and several workers suggested that the extremely lobate southern margin, together with chronological evidence of rapid re-advances, resulted from large scale surging (Wright, 1973; Clayton et al., 1985). Sustained ice streaming, rather than more temporary surge behaviour, was later suggested for several of the southern lobes (Patterson, 1997; Patterson, 1998; Jennings, 2006). Such behaviour was linked to the availability of fine-grained tills that generated low basal shear stresses (Hicock, 1988; Hicock et al., 1989; Hicock and Dreimanis, 1992). Alley (1991) suggested that these widespread till sheets were deposited as a deforming bed with ice velocities of hundreds m a^{-1}, similarly to that which had been proposed for the modern Ice Stream B (re-named Whillans Ice Stream) in West Antarctica (Alley et al., 1986; Alley et al., 1987). Indeed, based on these concepts and the known or assumed bed properties, Marshall et al. (1996) used numerical model to generate an ice stream likelihood map for the entire LIS, which further highlighted the north-western, western, and south-margins as being conducive to ice streaming because of the substrate.

As noted above, the late 1980s and 1990s, saw the discovery of layers of ice rafted debris (IRD) in marine sediment cores from the North Atlantic (Heinrich, 1988), which renewed interest in the behaviour of Hudson Strait Ice Stream: the anticipated source of icebergs carrying the terrestrial material found on the sea floor. Bond et al. (1992) identified the IRD material as originating from the region of Hudson Bay/Strait and episodes of increased calving from this region were constrained by the description and dating of individual Heinrich layers (Andrews and Tedesco, 1992; see Andrews, 1998, for a review). Conceptual models of these binge-purge oscillations were put forward, supported by numerical modelling experiments (e.g., the ‘Binge-Purge’ model: MacAyeal, 1993; Clark, 1994; Marshall and Clarke, 1997b).

The 1990s also saw the application and rapid expansion of remote sensing and Geographical Information Systems (GIS) techniques in palaeo-glaciology, which ushered in a new era of palaeo-ice stream research (Clark, 1993; Clark, 1997). This approach typically employs regional scale mapping of the glacial landform record to reconstruct past ice sheet dynamics, including ice streams (Kleman et al., 1997). Based largely on the characteristics of modern ice streams and the pioneering work by Dyke and colleagues (Dyke and Morris, 1988; Hicock, 1988; Dyke et al., 1992), criteria for the identification of palaeo-ice streams in the landform record were developed (Stokes and Clark, 1999; Stokes and Clark, 2001; Stokes, 2002). Subsequently, a number of individual ice stream tracks have been identified and examined (Clark and Stokes, 2001; Stokes and Clark, 2003a; Stokes and Clark, 2004; Stokes et al., 2005; Dyke, 2008; Ó Cofaigh et al., 2013b; Stokes et al., 2013) and regional reconstructions have been carried out that incorporate their temporal evolution (De Angelis and Kleman, 2005; De Angelis and Kleman, 2007; Evans et al., 2008; Ross et al., 2009; Stokes et al., 2009; Ó Cofaigh et al., 2010; Ross et al., 2011). These efforts have mostly focussed on the tundra regions of northern Canada, where sparse vegetation allows for easier landform recognition in satellite imagery (Fig. 5). More recent studies have successfully mapped portions of the Interior Plains using Digital Elevation Models (DEMs), despite intensive modification of the landscape due to agriculture and other human activity (Evans et al., 2008; Ross et al., 2009; Ó Cofaigh et al., 2010; Evans et al., 2014).

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Although terrestrial evidence and inferences had seen a large number of ice streams hypothesised in major marine channels (De Angelis and Kleman, 2005; Stokes et al., 2005; De Angelis and Kleman, 2007), there were only limited data on the morphology and stratigraphy of areas submerged by present-day sea-level. Recently, high-resolution swath bathymetry data have become available, albeit with limited extent in some areas, and studies by MacLean et al. (2010) and Ross et al. (2011) have made use of these data, describing sea-floor lineations from Hudson Bay, Franklin Strait, Peel Sound and M'Clintock Channel (Fig. 3). More extensive datasets are available from Atlantic Canada, which records a number of ice streams operating in glacial troughs carved into the continental shelf (Shaw, 2003; Shaw et al., 2006; Todd et al., 2007; Shaw et al., 2009; Todd and Shaw, 2012; Shaw et al., 2014). In their updated inventory, Margold et al. (in press) also used the International Bathymetric

Chart of the Arctic Ocean (IBCAO: Jakobsson et al., 2000) and more detailed swath bathymetry data from the Canadian Arctic (ArcticNet, 2013) to identify several new ice streams and confirm others that were previously hypothesised based only on terrestrial evidence. These bathymetric data have also been complemented by sub-surface data obtained from seismic reflection surveys, allowing workers to identify multiple till units, grounding zone wedges and other glacial features buried in the marine sediments; and to investigate the architecture of large trough mouth fans that often lie distal to the major ice stream troughs (Jennings, 1993; Andrews et al., 1995b; Rashid and Piper, 2007; Li et al., 2011; Siegel et al., 2012; Batchelor and Dowdeswell, 2014; Batchelor et al., 2013a; Batchelor et al., 2013b; Batchelor et al., 2014).

In addition to field and remote sensing studies, numerical modelling of the LIS has been used to explore ice stream activity in the LIS. One of the earliest studies was by Sugden (1977), who modelled the annual ice discharge of some of the largest ice streams/outlet glaciers. The activity of ice streams over deformable beds has also been replicated in numerical modelling experiments, especially at the southern margin (Fisher et al., 1985; Breemer et al., 2002; Winguth et al., 2004; Carlson et al., 2007; Meriano and Eyles, 2009). Topographically-controlled ice streams have also been modelled in areas of higher relief (Kaplan et al., 1999), and, as noted above, the oscillations of the Hudson Strait Ice Stream have attracted most attention, largely targetted at explaining Heinrich events (MacAyeal, 1993; Marshall et al., 1996; Marshall and Clarke, 1997a; Marshall and Clarke, 1997b; Calov et al., 2002). Pan-ice sheet models have also generated ice streams (e.g., Tarasov and Peltier, 2004), and a recent data-model comparison suggests that they are likely to capture most of the major ice streams, especially those that are topographically controlled (Stokes and Tarasov, 2010).

Finally, several workers have, periodically, attempted to summarise and inventorise the growing number of hypothesised ice streams. Patterson (1998) was one of the first to explicitly map the evidence for their location across the entire ice sheet, and this was updated by Winsborrow et al. (2004), who identified a total of 49 hypothesised locations, albeit with some more speculative than others. Extending this work, and building on several more recent studies and the burgeoning availability of sea-floor data, Margold et al. (in press) have compiled a new map of 117 ice streams. Margold et al. (in press) refrained from an in-depth analysis and discussion of the ice streams, which is the purpose of this paper.

3. Types of Evidence for Laurentide ice streams

Paterson (1994: p. 301) defined an ice stream as "*a region in a grounded ice sheet in which the ice flows much faster than in regions on either side*", which reiterates the original description by Swithinbank (1954). Although there has been some debate about what qualifies as an ice stream (Bentley, 1987; Bennett, 2003; Truffer and Echelmeyer, 2003), we follow this simple and concise definition in our review, i.e. it represents an abrupt spatial transition in ice-flow velocity (and which must be reflected in the evidence). This definition encompasses spatial transitions where an ice stream bordered by slower moving ice may then feed into an outlet glacier *sensu stricto*, which is bordered by rock-walls. However, it ignores the temporal aspect of the rapid ice flow, which has caused some confusion and conflation of ideas in the literature, especially in relation to land-terminating (terrestrial) ice streams, where the term is often used interchangeably with surging (Clayton et al., 1985; Patterson, 1997; Evans et al., 1999; Jennings, 2006; Evans et al., 2012). Generally, ice streaming is used to describe a sustained period of fast flow (decades to millennia), whereas surge-type glaciers exhibit a cycle of fast flow (typically years), followed by a quiescent phase that is of much longer duration (typically decades; Raymond, 1987). This should help differentiate surge-type behaviour from ice streaming, but we note that some modern-ice streams have been suggested to stagnate and reactivate (Bougamont et al., 2003; Hulbe and Fahnestock, 2007). This has also been suggested in the palaeo-record (Stokes et al., 2009) and some have even used the term ‘surging ice streams’ (Evans et al., 1999; Evans et al., 2012). In summary, we adhere to the simple definition of an ice stream as an abrupt spatial transition in flow, but place no constraints on the duration of flow.

Several different types of evidence have been used for identifying ice streams in the landform and sedimentary record (Stokes and Clark, 2001). In Margold et al.’s (in press) recent mapping inventory, these types of evidence are broadly categorised (see also Fig. 6) as:

(i) evidence of fast ice flow in the landform record – the ‘bedform imprint’ (Fig. 5)

(ii) evidence of glacial troughs (Fig. 7)

(iii) evidence of sedimentary depo-centres beyond the edge of the continental shelf (Fig. 7)

(iv) evidence of specific till characteristics suggested to be indicative of fast ice flow, or indicating a distinct sediment dispersal pattern

(v) ice rafted debris traced to its source region

In relation to (i), streamlined landforms such as drumlins, whalebacks and roches moutonnées have long been recognised as a product of basal sliding or sediment deformation under flowing ice (e.g., Boulton, 1987). Larger-scaled streamlined patterns in the form of mega-scale glacial lineations (MSGLs) have also been observed in satellite imagery (Fig. 5; Punkari, 1982; Boulton and Clark, 1990; Clark, 1993) and have been interpreted as a product of fast ice flow (Clark, 1993; Clark et al., 2003; Stokes et al., 2013). This interpretation has been confirmed by the observation of MSGLs under the Rutford Ice Stream in Antarctica (King et al., 2009) and a ridge-groove landform pattern under Jakobshavn Isbræ of the Greenland Ice Sheet (Jezek et al., 2011), with further support from landform assemblages on the beds of Greenland and Antarctic palaeo-ice streams (e.g. Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002; Wellner et al., 2006; Dowdeswell et al., 2008; Graham et al., 2009; Livingstone et al., 2012; Ó Cofaigh et al., 2013a) . As noted above, Stokes and Clark (1999) listed criteria for the identification of palaeo-ice streams defined by their bedform imprint (as opposed to those defined by large scale topography) and these are: characteristic shape and dimensions, highly convergent flow patterns (Fig. 4), highly attenuated bedforms (Fig. 5), abrupt lateral margins (Fig. 4), lateral shear margin moraines (Fig. 4), evidence of pervasively deformed till, Boothia-type dispersal trains (Fig. 4), and submarine till deltas or sediment fans. Not all of these criteria have to be present, but this is by far the most commonly utilised form of evidence (see Fig. 6). To account for the variable quantity and quality of evidence left behind by different ice streams, Margold et al. (in press) further sub-divided this type of landform evidence into three classes: (i) ice streams with full bedform imprint, (ii) ice streams with discontinuous bedform imprint, and (iii) ice streams with isolated bedform imprint. For the last group, if no other evidence has been found to constrain the ice stream extent, then it is described as an ice stream fragment (Margold et al., in press).

***Fig. 6 here**, column width*

***Fig. 7 here**, column width*

In relation to type (ii), some LIS ice streams have been inferred almost exclusively from large-scale topography (e.g., the Massey Sound Ice Stream; Fig. 3; England et al., 2006). To search for and identify this type of evidence, Margold et al. (in press) mapped prominent glacial troughs across the entire LIS bed, both onshore and offshore, which resulted in a number of newly-identified ice streams. This mapping also included the identification of type (iii) evidence in the form of sedimentary depo-centres beyond the edge of the continental shelf (expressed in the form of a contour bulge at the shelf edge in the topographic data) and benefitted from similar surveys undertaken for the entire Arctic (Batchelor and Dowdeswell, 2014).

Type (iv) evidence (sedimentological) is usually reported in conjunction with type (i) evidence (Fig 6; Kehew et al., 2005; Ross et al., 2011), but has only been reported for a handful of ice streams, compared to type (i) evidence. Similarly, type (v) evidence (IRD) has perhaps been under-utilised in the literature, but can be a powerful constraint on the timing of ice stream operation (Stokes et al., 2005; Rashid et al., 2012).

Clearly, the robustness of evidence varies widely among the identified ice streams (Fig. 6). Whereas some ice streams are hypothesised based on a variety of different lines of evidence (e.g., the Cumberland Sound, Amundsen Gulf, or M'Clure Strait ice streams) others are inferred only from one type of evidence and their existence is therefore more speculative (e.g., the Rocky Mountain Foothills, Quinn Lake, or offshore Massachusetts ice streams; Fig. 6; Supplementary data). It is also possible that some ice streams operated but left behind very little (if any) evidence, and we discuss the possibility of ice streams being missed in Section 5.1.

4. An Updated Inventory of Laurentide Ice Streams

In this section, we provide a brief review of the location and operation of ice streams from different sectors of the LIS (see Margold et al., in press). We do this according to five major physiographic regions, which likely influenced the pattern of ice dynamics. These are: (1) the Canadian Arctic Archipelago, (2) the Interior Plains, (3) the Great Lakes, (4) the Atlantic seaboard, and (5) the Canadian Shield (Fig. 2). Note that detailed information about the evidence used to identify each ice stream is available in the Supplementary data accompanying this paper.

4.1. Canadian Arctic Archipelago

The islands of the Canadian Arctic Archipelago (CAA) are built largely from sedimentary rocks, except in the east where the SE part of Ellesmere Island and much of Baffin Island consist of crystalline rocks (Fig. 8). The depth of the channels between the islands is generally not greater than 500 m, but deeper areas (up to 1100 m) can occur, many of which exhibit characteristics of glacial overdeepenings (Cook and Swift, 2012), such as Nansen Sound, Jones Sound, Smith Sound, Robeson Channel or Lancaster Sound in the north of the archipelago, as well as Cumberland Sound and parts of Hudson Strait (Fig. 3).

***Fig. 8 here**, column width*

The northernmost part of the CAA hosted an independent ice mass, the Innuitian Ice Sheet, which was confluent with the LIS during glacial maxima and connected by a saddle above Nares Strait to the Greenland Ice Sheet (Fig. 2; Funder and Hansen, 1996; Dyke, 1999; England, 1999; England et al., 2006). The saddle was drained by ice streams to the north, through Robeson Channel (no. 141 in Fig. 3; Jakobsson et al., 2014; Margold et al., in press), and to the south, through Smith Sound (no. 126 in Fig. 3; Blake et al., 1996; England, 1999;

England et al., 2004, 2006; Margold et al., in press; Simon et al., 2014), where distinct glacial lineations appear in swath bathymetry data (Supplementary data). Three relatively small, glacially eroded troughs occur on the shelf off the NW coast of Ellesmere Island, with only the northernmost of these crossing the whole shelf and forming a pronounced sediment bulge at the shelf-break (nos. 125, 139, and 140 in Fig. 3; Margold et al., in press). A larger ice stream has been inferred in Nansen Sound that cuts into the central parts of Ellesmere Island, where it forms a large, branching, overdeepened fjord (no. 124 in Fig. 3; Sugden, 1977; Bednarski, 1998; England et al., 2006; Jakobsson et al., 2014; Margold et al., in press). Distinct lateral ridges border the trough on the outer shelf, protruding beyond the shelf edge (Margold et al., in press). Two relatively extensive, broad ice streams have been suggested to drain the southern part of the Innuitian Ice Sheet to the northwest (nos. 123 and 129 in Fig. 3; Lamoureux and England, 2000; Atkinson, 2003; England et al., 2006; Jakobsson et al., 2014; Margold et al., in press). These inferences have been based on the topography on the shelf, with the northern one in Massey Sound bordered by lateral ridges, forming a gentle bulge in the shape of the shelf edge (Margold et al., in press).

The south-western region of the CAA hosted two major ice streams that operated during the LGM and deglaciation (nos. 18 and 19 in Fig. 3), both occupying major channels – M'Clure Strait and Amundsen Gulf – and draining a large portion of the Keewatin Ice Dome (Figs. 2, 3). Both ice streams formed a distinct trough mouth fan beyond the edge of the continental shelf, and their sedimentary record contains grounding zone wedges close to the shelf edge (Batchelor et al., 2013a; Batchelor et al., 2013b; Batchelor et al., 2014). The swath bathymetry data from M'Clure Strait Ice Stream are dominated by iceberg scours (Supplementary data). The main evidence for the ice stream comes from the cross-shelf trough and trough mouth fan, together with terrestrial landform record on the surrounding islands, where several ice stream flow-sets have been identified (Hodgson, 1994; Stokes et al., 2005; England et al., 2009; Stokes et al., 2009). The M'Clure Strait Ice Stream is thought to have operated episodically during deglaciation, with a shorter M'Clintock Channel Ice Stream operating prior to final deglaciation (no. 10 in Fig. 3; Clark and Stokes, 2001; Stokes, 2002; De Angelis and Kleman, 2005; Stokes et al., 2005; Stokes et al., 2009; MacLean et al., 2010). In contrast, the Amundsen Gulf Ice Stream is thought to have operated throughout deglaciation, and was spatially more stable than both the M'Clure Strait Ice Stream to the north and the Mackenzie Trough Ice Stream to the west (Stokes et al., 2009; Brown, 2012). It is evidenced both by terrestrial landform record on the mainland and on Victoria Island (Sharpe, 1988; Stokes et al., 2006; Kleman and Glasser, 2007; Storrar and Stokes, 2007; Stokes et al., 2009; Brown et al., 2011; Brown, 2012), and by erosion and distinctly streamlined morphology of the seabed in Amundsen Gulf (Supplementary data; Batchelor et al., 2013b).

During deglaciation, a number of smaller ice streams (50-150 km long, 10-50 km wide) also operated on Victoria and Prince of Wales islands, in or near the catchments of the M'Clure Strait/M'Clintock Channel and Amundsen Gulf ice streams, and mostly resulting from the opening up of major marine embayments (nos. 7, 8, 11, 12, 101, 102 in Fig. 3; De Angelis and Kleman, 2005; Stokes et al., 2005; Stokes et al., 2009).

In addition to the ice streams draining Keewatin ice towards the Beaufort Sea, they also existed in Peel Sound and the Gulf of Boothia (nos. 13 and 20 in Fig. 3; Dyke and Dredge, 1989; Dredge, 2000, 2001; De Angelis and Kleman, 2005, 2007; Kleman and Glasser, 2007; MacLean et al., 2010), draining Keewatin ice to the north, where it was captured by the W-E oriented Lancaster Sound (no. 22 in Fig. 3; De Angelis and Kleman, 2005; Briner et al., 2006). The Gulf of Boothia Ice Stream also drained ice from the Foxe Ice Dome across the Melville Peninsula and around Baffin Island (Figs. 2, 3). The major trunk ice stream in Lancaster Sound has also been suggested to have been joined from the north by a tributary in Wellington Channel draining Innuitian ice (no. 128 in Fig. 3; Fig. 2; Dyke, 1999; England et al., 2006), although there is little evidence for this tributary ice stream. In contrast, another tributary ice stream in Jones Sound (no. 127 in Fig. 3), joining the Lancaster Sound Ice Stream from the north on the continental shelf in the north-western part of Baffin Bay, has left a distinctly streamlined bed visible in the swath bathymetry data (Supplementary data).

The Lancaster Sound Ice Stream, draining Keewatin, Foxe and Innuitian ice, formed one of the major arteries in the NE sector of the LIS (Figs. 2, 3; Sugden, 1977; De Angelis and Kleman, 2005; Briner et al., 2006; Simon et al., 2014), which is evidenced by a major sediment bulge protruding into Baffin Bay (Li et al., 2011; Batchelor and Dowdeswell, 2014). The divide between the M'Clure Strait and Amundsen Gulf Ice Stream catchments and the Lancaster Sound Ice Stream catchment was probably highly mobile, and there is evidence for ice piracy whereby Keewatin ice was captured from the onset zone of the M'Clure Strait (later M'Clintock Channel) Ice Stream across the Boothia Peninsula and southern Somerset Island into the Lancaster Sound Ice Stream catchment (Figs. 2, 3; De Angelis and Kleman, 2005).

Apart from the drainage around Baffin Island through the Gulf of Boothia, Foxe ice also drained across Baffin Island. Two major routes in the NW of Baffin Island were Admiralty Inlet and Eclipse Sound (nos. 21 and 104 in Fig. 3; De Angelis and Kleman, 2007). In the central parts of the island, ice was funnelled through narrow fjords, a product of selective linear erosion across many glacial cycles (Løken and Hodgson, 1971; Sugden, 1978), with ice from several fjords typically feeding one cross-shelf trough (nos. 108-116, and 172 in Fig. 3; Fig. 7; Løken and Hodgson, 1971; Briner et al., 2006; De Angelis and Kleman, 2007; Briner et al., 2008; Batchelor and Dowdeswell, 2014; Margold et al., in press). From the east, Foxe ice was also drained across SE Baffin Island by two sizable ice streams in Cumberland Sound and Frobisher Bay (nos. 23 and 117 in Fig. 3; Sugden, 1977; Kaplan et al., 2001; Andrews and MacLean, 2003; Briner et al., 2006; De Angelis and Kleman, 2007). The Foxe ice drainage pattern appears to have been relatively stable during the LGM, and in the early stages of the ice sheet retreat, but changed dramatically during the collapse of the Foxe Ice Dome when a number of small, ephemeral deglacial ice streams operated on Baffin Island, with ice flow directions often opposite to those at the LGM (nos. 103, 106, 107, 118-120 in Fig. 3; De Angelis and Kleman, 2007, 2008; Dyke, 2008).

The Hudson Strait Ice Stream was one of the largest in the LIS and is probably the most studied (no. 24 in Fig. 3; Supplementary data). Its onset zone was in the vicinity of Hudson Bay and ice was routed through Hudson Strait to the shelf of the Labrador Sea (Figs. 2, 3; Andrews and MacLean, 2003; De Angelis and Kleman, 2007; Rashid and Piper, 2007; Ross et al., 2011). It drained the central parts of the ice sheet, receiving ice from all the major domes: Keewatin, Foxe and Labrador (Figs. 2, 3). However, the evidence of ice streaming is actually rather sparse, compared to some other ice streams with a fuller bedform imprint, and mainly comprises long-distance erratic dispersal to the shelf and IRD of Hudson Bay provenance (Andrews and MacLean, 2003; Rashid and Piper, 2007; Rashid et al., 2012). The landform record is not always obvious (Hulbe et al., 2004; De Angelis and Kleman, 2007), but Ross et al. (2011) described a streamlined zone west of Hudson Bay as a possible onset zone of the Hudson Strait Ice Stream. Ice stream flow-sets on Southampton Island (nos. 121 and 122) have been interpreted to postdate the period of the Hudson Strait Ice Stream operation and originate from later deglacial ice streams (Fig. 3; De Angelis and Kleman, 2007; Ross et al., 2011).

Whilst the identification of IRD layers in the North Atlantic with a high detrital carbonate content (Heinrich, 1988) has been linked to the activity of the Hudson Strait Ice Stream, little is known about the response of other ice streams along the Atlantic seaboard, or farther afield (see Mooers and Lehr, 1997; Dyke et al., 2002; Stokes et al., 2005). However, Heinrich events 5, 4, 2, and 1 appear to originate from the Hudson Bay area (with H4 being the strongest), whereas H6, H3, and H0 are more likely of Ungava origin with H6 and H3 also having a large contribution of European sources (Hemming, 2004).

In summary, the CAA exhibits robust evidence of numerous ice streams draining the major ice domes towards marine margins and in a pattern that is not entirely dissimilar to the present-day situation in West Antarctica (cf. Figs. 1 and 2). Ice streaming in the region of the CAA was concentrated in large, broad, marine channels where weaker sedimentary rocks and unconsolidated marine sediments enhanced fast ice flow. In contrast, the fjord landscapes along the coasts of Baffin and Ellesmere islands were more analogous to the high relief coasts of Greenland and East Antarctica (e.g., Dronning Maud Land; Fig. 7). The timing of ice stream activity has been studied only in the south-western part of the CAA and in association with the Hudson Strait Ice Stream and its role during Heinrich events.

4.2. Interior Plains

The western margin of the LIS extended into the region of the Interior Plains, an area of low relief built predominantly of soft sedimentary rocks (Fig. 8; Fig. 9). A number of ice streams have been identified in this area, although it has received relatively little attention and is one of the less well-understood sectors of the ice sheet. In the northwest, a large drainage system existed along the course of the present-day Mackenzie River, but it may have reached the continental shelf on fewer occasions than the ice streams further north and east in Amundsen Gulf and M'Clure Strait (Batchelor et al., 2013a; Batchelor et al., 2013b). The shallow Mackenzie Trough appears to have formed the main ice discharge route, but the landform record indicates that ice drainage was highly dynamic and ice streams operated along different trajectories (Fig. 9; Brown et al., 2011; Brown, 2012; Margold et al., in press). Tracks of four major ice streams have been reconstructed in the area: The Mackenzie Trough, Anderson, Bear Lake, and Fort Simpson ice streams (nos. 1, 2, 5, and 144 in Fig. 9; Brown, 2012; Batchelor et al., 2014; Margold et al., in press). However, it is not entirely clear whether these were separate ice streams or different trajectories of a major ice stream system changing its course over time (Brown, 2012). East of the major ice streams of the Mackenzie region, three smaller, well-defined ice streams developed during ice retreat: the Horton/Paulatuk, Haldane, and Kugluktuk ice streams (nos. 3, 4, and 142 in Fig. 9; Winsborrow et al., 2004; Kleman and Glasser, 2007; Brown, 2012; Margold et al., in press).

***Fig. 9 here**, column width*

An area of coalescence of the LIS with the Cordilleran Ice Sheet (CIS) existed during the LGM between 62° and 52° N and this saddle provided ice that drained through the Mackenzie region to the north (Figs. 2, 9). Several troughs with generally westerly orientation are also found near this saddle area in SW Northwest Territories and in N Alberta, between the higher plateau surfaces of the Cameron Hills, Caribou Mountains and Birch Mountains (Figs. 9). The landform record is patchy in this region (see Fig. 10) and ice drainage has not been studied in detail. However, Margold et al. (in press) have mapped topographically inferred ice streams draining to the west through these troughs (nos. 175-178 in Fig. 9). Fragmented evidence of fast ice flow has also been found on the plateau surfaces of the Cameron Hills and the Birch Mountains (nos. 145 and 148 in Figs. 9, 10; Margold et al., in press), indicating a period of fast ice flow unconstrained by the regional topography. These ice streams draining to the west could have operated before the CIS and LIS coalesced, or their operation could have again commenced after the CIS-LIS ice saddle collapsed rapidly during deglaciation (see Gregoire et al., 2012), followed by topographically constrained ice streams.

***Fig. 10 here**, column width*

The south-western Interior Plains, in Alberta and Saskatchewan, exhibit one of the most complex networks of ice stream activity documented for a Northern Hemisphere Pleistocene ice sheet (Figs. 2, 9). Ice stream trajectories in this region have orientations varying from WSW to SE, most probably indicating an evolving ice stream network during the ice sheet advance (ice flow to WSW) , maximum extent (ice flow to SW and S) and retreat (ice flow changing from S to SE and then back to S and finally SW;Fig. 9; Evans, 2000; Evans et al., 2008; Ross et al., 2009; Ó Cofaigh et al., 2010; Evans et al., 2012, 2014; Margold et al., in press). The complex network of streamlined corridors has also been interpreted as reflecting the paths of subglacial mega-floods (e.g., Shaw, 1983; Rains et al., 1993; Shaw et al., 1996, 2000; 2010), rather than ice streams. This interpretation has been the subject of much debate (e.g., Benn and Evans, 2006; Evans, 2010; Shaw, 2010a, b; Evans et al., 2013; Shaw, 2013), which is yet to be fully resolved, not least because ice streams are typically associated with abundant subglacial meltwater that helps lubricate their flow. However, questions remain, for example, regarding the sources and volume of water required to feed putative mega-flood tracks (Clarke et al., 2005). Thus, spatially-confined fast-flowing ice (ice streaming) is the simpler interpretation at present and one which we adopt in this paper.

Two long and narrow ice stream tracks run across central Alberta in N-S direction: the Central Alberta Ice Stream and the High Plains Ice Stream (nos. 14 and 15 in Fig. 9; Evans, 2000; Evans et al., 2008; Ross et al., 2009; Evans et al., 2012, 2014), and a more complicated network of ice streams occurs further east in Saskatchewan (Fig. 9; Ross et al., 2009; Ó Cofaigh et al., 2010; Evans et al., 2014; Margold et al., in press). To the south of the complex flow record in central Saskatchewan, two major ice lobes protruded from the southern LIS margin: the James Lobe and the Des Moines Lobe (nos. 28 and 27 in Fig. 9). Both are thought to be formed by ice streams operating during the Late Glacial, draining ice from Saskatchewan and Manitoba along the ice sheet margin in a SSE direction to North and South Dakota, Minnesota and Iowa (Clayton et al., 1985; Clark, 1992; Patterson, 1997; Jennings, 2006; Lusardi et al., 2011). After a considerable retreat of the ice margin, ice streaming (surging in Clayton et al., 1985; Dredge and Cowan, 1989) is documented in a southerly direction for the late stage of the Red River Lobe (Margold et al., in press).

In summary, the Interior Plains contain evidence of numerous ice streams which drained ice northwards to the Beaufort Sea coast, westward to the Rocky Mountains, and south-westward and south-eastward, towards the southern margin of the ice sheet (Fig. 9). Ice streaming on the Interior Plains was enhanced by the presence of weak sedimentary bedrock and occurred in broad, shallow troughs creating sinuous corridors of smoothed terrain (controls on ice stream location are discussed in depth in Section 5.4). However, the ice sheet geometry that defined the pattern of ice drainage is poorly understood, especially in relation to the pattern and timing of the CIS-LIS coalescence and its collapse.

4.3. Great Lakes

The Great Lakes basins developed under recurring glaciations by glacial erosion of river valleys in a region of relatively weak sedimentary rocks (Fig. 8; Larson and Schaetzl, 2001). The deepest basin is Lake Superior, which has a floor at an elevation of 213 m below sea level and a depth of almost 400 m measured from its outlet (Larson and Schaetzl, 2001). During the last glaciation, the basins of the Great Lakes constituted a major topographic control on ice flow, which resulted in a lobate ice margin during the LGM and during ice retreat (Karrow, 1989; Mickelson and Colgan, 2003). The maximum extent during the Late Wisconsinan was attained earlier than in the James and Des Moines Lobes to the west (Mickelson et al., 1983; Hallberg and Kemmis, 1986; Mickelson and Colgan, 2003). This was most apparent at the contact between the Des Moines Lobe and the Superior Lobe, where the latter retreated to the NE and the Grantsburg Sub-lobe of the Des Moines Lobe advanced into the area formerly occupied by the Superior Lobe (Figs. 9, 11; Jennings, 2006).

***Fig. 11 here**, column width*

The dominance of the particular lobes in the Great Lakes region changed through time (Mickelson and Colgan, 2003; Kehew et al., 2005). The most extensive was an advance of the Saginaw Lobe over the southern Michigan Upland (Fig. 11; Kehew et al., 2005) and ice streaming has been inferred for the area where the lobe passed over Huron Lake (no. 184 in Fig. 11; Eyles, 2012). When the Saginaw Lobe retreated, ice advanced in the surrounding areas through the basins of Lake Michigan and lakes Ontario and Erie (Fig. 11; Dyke et al., 2003; Kehew et al., 2005). The Huron-Erie Lobe Ice Stream, occupying the basins of Lake Ontario and Lake Erie, received ice that previously fed the Saginaw Lobe and was now diverted through the basin of Lake Huron to the south instead of flowing through the Saginaw

Bay to the SW (no. 49 in Fig. 11; Kehew et al., 2005). In addition to topographic steering, fine lacustrine sediments were conducive to fast ice flow (Beget, 1986; Clark, 1992; Hicock, 1992; Hicock and Dreimanis, 1992; Lian et al., 2003; Kehew et al., 2005; Jennings, 2006; Eyles, 2012; Kehew et al., 2012; discussed further in Section 5.4.3). Rapid ice flow is further supported by observations of glacial lineations on the floor of Lake Ontario and by glacial landsystems composed of drumlin fields, tunnel valleys, thrust blocks, and recessional moraines (Jennings, 2006; Eyles, 2012; Kehew et al., 2012). A more localised ice stream has also been reconstructed for the Oneida Lobe and the higher ground of the Tug Hill Plateau in the region east of Lake Ontario (nos. 136 and 137 in Fig. 11; Briner, 2007; Margold et al., in press).

In summary, several large ice streams have been active in the basins of the Great Lakes. Ice streaming in this region has been inferred from the wide-scale topography and from landsystems identified to be characteristic of fast ice flow (Kehew et al., 2005; Jennings, 2006; Hess and Briner, 2009; Eyles, 2012; Kehew et al., 2012; Margold et al., in press).

4.4. Atlantic seaboard

The Atlantic seaboard of North America hosts the broadest section of the continental shelf covered by the LIS (Fig. 2). The region is built by crystalline lithologies of the Canadian Shield landwards from the coast-parallel Marginal Trough on the NE Labrador coast, down to the coast of the Gulf of St Lawrence in SE Labrador, and in most of Newfoundland (Figs. 8, 12). Sedimentary lithologies occur on most of the continental shelf and in the Northern Appalachians (Fig. 8).

Fig. 12 here*, column width*

The eastern margin of the LIS featured a number of ice streams crossing the present-day continental shelf (Figs. 2, 12; Margold et al., in press). Only limited evidence exists for ice streams in the Gulf of Maine, where the most prominent feature is the trough of the Northeast Channel Ice Stream (no. 134 in Fig. 12; Shaw et al., 2006). Another major ice-discharge route for Labrador ice constituted the Laurentian Channel Ice Stream (no. 25 in Fig. 12; Grant, 1989; Keigwin and Jones, 1995; Shaw et al., 2006, 2009; Eyles and Putkinen, 2014). This ice stream occupied a well-defined trough that runs for more than 700 km from the Gulf of St Lawrence to the shelf edge, and with an overdeepening of about 400 m and a width of 70 to 100 km. As such, a deep calving bay has been inferred to have developed in the Gulf of St Lawrence during deglaciation, forcing a significant retreat of the Laurentian Channel Ice Stream at the time when ice complexes still survived on Newfoundland and Nova Scotia, and drained to the ocean through several smaller ice streams (Stea et al., 1998; Shaw, 2003; Shaw et al., 2006; Todd et al., 2007; Shaw et al., 2009; Todd and Shaw, 2012). Indeed, Newfoundland hosted an independent ice complex that was drained to the north and north-east by ice streams in the Notre Dame Channel and the Trinity Trough (nos. 45 and 130 in Fig. 12; Shaw, 2003; Shaw et al., 2006, 2009; Rashid et al., 2012), and which fed into the Laurentian Channel Ice Stream to the south. Ice from Newfoundland was also drained through the Placentia Bay-Halibut Channel Ice Stream (no. 133 in Fig. 12; Shaw, 2003; Shaw et al., 2006). Prominent troughs also occur off the NE Labrador coast, most of them reaching the shelf edge. Although the subject of relatively little research, they are likely to have hosted palaeo-ice streams draining the Labrador Ice Dome (nos. 167-171 in Fig. 12; Fig. 2; Josenhans et al., 1986; Josenhans and Zevenhuizen, 1989; Rashid et al., 2012; Margold et al., in press).

In summary, the Atlantic seaboard exhibits strong evidence for focused drainage of Labrador ice in a number of ice streams that incised distinct troughs in the continental shelf. The region centred on the Gulf of St Lawrence has been the subject of a series of studies documenting the role of ice streams during deglaciation (Shaw, 2003; Shaw et al., 2006; Todd et al., 2007; Shaw et al., 2009; Todd and Shaw, 2012), but the NE Labrador coast and the adjacent shelf have received comparatively less attention.

4.5. Canadian Shield

The Canadian Shield formed the interior of the LIS during its maximum extent and hosted two of the three major ice domes: Keewatin and Labrador (Figs. 2, 13). It is built of crystalline lithologies and its landscapes are characterised by low relief with a dominance of areal scouring (Figs. 8, 14; Sugden, 1978; Krabbendam and Bradwell, 2014).

***Fig. 13 here**, full page width*

***Fig. 14 here**, full page width*

It was only after substantial ice retreat (by 12 ka BP) that ice margins were located over the Shield (Dyke and Prest, 1987a; Dyke et al., 2003) and a number of ice streams have been hypothesised in this smaller, deglaciating LIS (Fig. 2). Arguably, the best studied of the deglacial ice streams of the Canadian Shield is the NW-flowing Dubawnt Lake Ice Stream in northern Keewatin (no. 6 in Fig. 13). This large broad ice stream has been reconstructed from its distinct bedform imprint, which is one of the best preserved on the entire ice sheet bed (Fig. 5; Kleman and Borgström, 1996; Stokes and Clark, 2003a, b, 2004; De Angelis and Kleman, 2008; Ó Cofaigh et al., 2013b; Stokes et al., 2013). Other major ice streams formed at the south-western margin of the retreating ice, such as the Hayes Lobe and the Rainy Lobe (nos. 179 and 180 in Fig. 13; Dredge and Cowan, 1989; Margold et al., in press). Although the fan-shaped tracks of these ice streams are atypical, both of these large lobes fulfil the definition of an ice stream as a spatially defined partitioning of ice flow (see Section 3.).

East of the Rainy Lobe, ice was drained by a succession of ice streams, with the Albany Bay Ice Stream initially operating along the trajectory stretching from James Bay along the Albany River to the Lake Superior basin (no. 26 in Fig. 13; Hicock, 1988), and followed by the James Bay Ice Stream that occupied James Bay and flowed in a southerly direction (no. 33 in Fig. 13; Parent et al., 1995; Veillette, 1997). Apart from the Dubawnt Lake Ice Stream, none of the ice streams around Hudson Bay has received detailed scrutiny. The Quinn Lake Ice Stream (no. 164 in Fig. 13), mapped by Margold et al. (in press) is depicted as a distinct local readvance in the map of Dyke and Prest (1987b) whereas the Ekwan River Ice Stream is there portrayed as a series of minor lobes (no. 165 in Fig. 13; Dyke and Prest, 1987b). The Ekwan River Ice Stream was later identified by Thorleifson and Kristjansson (1993). We speculate on the nature of the unusually broad ice streams of the Canadian Shield in Section 5.2.

Relatively few ice streams have been reconstructed during deglaciation of the Labrador Ice Dome, especially at the SE and NE margins, after they had retreated from the shelf (Dyke and Prest, 1987b, c; Dyke et al., 2003). The most distinct features draining the Labrador Dome were a series of ice streams that drained ice in a northerly direction towards Ungava Bay (nos. 16, 17, and 188 in Fig. 13; Veillette et al., 1999; Clark et al., 2000; Jansson et al., 2003; Margold et al., in press). It is yet to be resolved whether they correlate with the putative H-0 event (11-10.5 ^{14}C ka, i.e. during the Younger Dryas; Andrews et al., 1995a; Andrews and MacLean, 2003) or with the Gold Cove and Noble Inlet advances (9.9-9.6 resp. 8.9-8.4 ^{14}C ka; Miller et al., 1988; Stravers et al., 1992; Kaufman et al., 1993; Jennings et al., 1998; Kleman et al., 2001) when the Labrador ice flowed across Hudson Strait in a NE direction.

In summary, conditions for ice streams on the Canadian Shield differed from other regions of the LIS: ice streams were not constrained by topography across these low-relief landscapes, and there were fewer fine-grained sediments available to lubricate their flow. They also operated late into the deglaciation and, as such, drained a much smaller ice sheet in a much warmer climate. Nevertheless, the region still supported large fan-shaped flow-sets that fit the definition of ice streams as spatially partitioned ice flow.

5. Discussion

5.1. To what extent have all of the LIS ice streams been found?

Our knowledge of palaeo-ice streams has grown rapidly in the last two decades (e.g., Stokes and Clark, 2001; Stoker and Bradwell, 2005; Andreassen et al., 2008; Winsborrow et al., 2010; Livingstone et al., 2012; Winsborrow et al., 2012; Roberts et al., 2013) and the LIS has played a key role in this regard (e.g., Patterson, 1997, 1998; De Angelis and Kleman, 2005, 2007, 2008; Ross et al., 2009; Stokes et al., 2009). Indeed, it now has the most comprehensive ice stream inventory of any of the former mid-latitude palaeo-ice sheets (Margold et al., in press), but an obvious question to ask is: are any ice stream locations missing?

The vast majority of hypothesised ice streams are informed by distinct bedform imprints (Fig. 6). These imprints are intimately linked to the availability of unconsolidated sediments that are moulded into a distinctive geomorphological signature (cf. Stokes and Clark, 1999) by the mechanisms that generate fast ice flow. However, there has been much debate about the processes that facilitate the fast flow of ice streams – whether through pervasive deformation of a metres thick layer of sediments at the bed or through basal sliding and/or with only a relatively thin layer of shearing at the top or within the sediments (Alley et al., 1986; Blankenship et al., 1986; Alley et al., 1987; Engelhardt et al., 1990; Engelhardt and Kamb, 1998). In the LIS, this has been particularly important for interpretations of the landform record associated with the southern ice lobes/streams (Clayton et al., 1985, 1989; Alley, 1991; Clark, 1991, 1992; Piotrowski et al., 2001; Hooyer and Iverson, 2002). Indeed, resolving this issue has implications for the identification of palaeo-ice streams and for wider inferences about long-term landscape modification by glaciers and ice sheets, because different flow mechanisms may modify the underlying landscape to a different degree.

Recent observations support the existence of both basal sliding and sediment deformation at the bed, which is best described by a plastic rather than viscous rheology (Iverson et al., 1995; Tulaczyk, 2006; Iverson et al., 2007; Smith and Murray, 2009; Reinardy et al., 2011). Furthermore, our ability to image the geomorphology at the bed of active ice streams has increased our confidence to identify them in the palaeo-record, confirming that mega-scale glacial lineations form under ice streams in areas of 'soft', deformable sediment (Smith et al., 2007; King et al., 2009). However, in cases where ice streaming might be facilitated only by sliding on a film of water and/or over more rigid (i.e. bedrock) surfaces, one might ask: what form of evidence does rapid sliding leave behind and how might we distinguish palaeo-ice streams in these settings? More generally, how are large volumes of sediment entrained and transported in these settings and what processes erode deep troughs?

Basal sliding across hard bedrock or within a shallow layer of underlying sediments (e.g., 3-25 cm: see Engelhardt and Kamb, 1998) might leave little evidence in the geological record and there are large areas of the LIS bed that are flat and without substantial thickness of sediment (e.g., the Canadian Shield). Theoretically, fast ice flow could have been facilitated by high subglacial water pressures that decoupled the ice from the bed (e.g., Zwally et al., 2002). Indeed, such 'hard-bedded' ice streams (i.e. spatially discrete fast ice flow over less erodible and mostly crystalline bedrock with little or no sediment cover) have been discussed for the Pleistocene Greenland Ice Sheet in central West Greenland (Roberts and Long, 2005; Roberts et al., 2010, 2013), the Fennoscandian Ice Sheet in south-western Finland (Punkari, 1995) and the British-Irish Ice Sheet in Scotland, where large mega-grooves have been interpreted to result from fast ice flow (Bradwell, 2005; Bradwell et al., 2008; Krabbendam and Glasser, 2011; Krabbendam and Bradwell, 2014). Interestingly, similar ridge-groove structures have recently been imaged beneath Jakobshavn Isbræ in West Greenland (Jezek et al., 2011). Recent work by Eyles (2012) and Eyles and Putkinen (2014) has also described rock drumlins, megaflutes and mega-lineated terrain, and argued that these landscapes represent a hard-bedded landform assemblage cut by ice streams. Indeed, even in hard bedrock terrains, there can be evidence of faint streamlined patterns visible in satellite imagery. For example, such regions exist around the Rae Isthmus in northern Keewatin (Fig. 3) and across parts of Baffin Island. De Angelis and Kleman (2007) interpreted these to represent small deglacial ice streams in areas of scoured bedrock around Amadjuak Lake on Baffin Island (Fig. 3), whereas the area of the Rae Isthmus has been interpreted as an onset zone of the Gulf of Boothia Ice Stream (Fig. 3; De Angelis and Kleman, 2007). Elsewhere, even when subglacial bedforms were not generated, there are zones of spatially discrete streamlined terrain that exhibit a smoothness not seen in the surrounding landscape. These are most obvious in the Interior Plains (Section 4.2.) and, in this context, Patterson (1998) noted that the finer the fraction composing the till, the fewer streamlined landforms are developed.

Apart from hard-bedded ice streams in heavily scoured bedrock zones and evidence of smooth ice stream tracks in the Interior Plains, there are other regions with wide-spread streamlining of predominantly bedrock terrain, but with thin veneers of sediment, particularly in NE Keewatin (Shaw et al., 2010; Kleman, unpublished). Whilst the degree of bedform attenuation and the general character of streamlined landscape indicate fast ice flow over thin veneers, the lateral boundaries of some of these zones are often extremely indistinct and preclude their classification as ice streams. Even the well-studied Dubawnt Lake Ice Stream (no. 6 in Fig. 13; Stokes and Clark, 2003a, b) has a rather 'blurred' northern margin. Thus, we cannot rule out the possibility that short-lived episodes of fast flow qualifying as ice streams have passed unnoticed in regions of extensive predominantly bedrock terrain, largely because our criteria for mapping ice stream tracks from remotely-sensed data (e.g., Stokes and Clark, 1999) do not account for hard-bedded ice streams, although there is potential to develop them (see Roberts and Long, 2005; Eyles, 2012; Eyles and Putkinen, 2014).

To conclude, we would argue that, as a result of more than 30 years of research, no large/major ice streams have been missed for the LIS, especially as Margold et al. (in press) specifically searched across the whole ice sheet bed in both onshore and offshore areas. That said, there remain some sectors of the LIS that are still poorly understood (e.g., the western-most margin), and other regions exist where hard-bedded, possibly short-lived deglacial ice streams may have existed but have not been reliably reconstructed.

5.2. Size and shape and comparison to modern ice streams

Present-day and palaeo-ice streams span across a wide range of sizes with lengths from tens to hundreds of km and widths of hundreds of metres to more than a hundred km (Figs. 1, 2; Rignot et al., 2011b; Margold et al., in press). It is important to note, however, that reconstructed tracks of palaeo-ice streams may not represent the extent of an ice stream at a particular point in time but rather the cumulative effect of evolving ice stream trajectories (cf. De Angelis and Kleman, 2005; Kleman and Glasser, 2007). This is likely to apply to ice streams that are not strongly controlled by topography, but even ice streams confined to deep troughs may have evolved with respect to the size of catchment they drained, thereby affecting the shape and size of their onset zone, the vigour of their flow, and their overall length (De Angelis and Kleman, 2005). Furthermore, although criteria have been defined to distinguish between time-transgressive and isochronous ice streams (and hence varying lengths of an ice stream within the same track; Stokes and Clark, 1999) our knowledge of the operational length of palaeo-ice streams is sufficient only for those with distinguishable onset zones preserved in the palaeo-record (De Angelis and Kleman, 2008). Indeed, contemporary velocity datasets often show a gradual and diffuse transition from slow-moving interior ice into more rapidly flowing ice streams (Bamber et al., 2000; Rignot et al., 2011b).

Notwithstanding the subjectivity in identifying when an ice stream actually ‘starts’ in the spatial sense, present-day Antarctic and Greenland ice streams display a variety of shapes. Most commonly, modern ice streams exhibit a dendritic pattern where the main trunk is fed by several tributaries (Fig. 1; Joughin et al., 1999; Rignot et al., 2011b). While some ice streams have long sinuous tributaries (e.g., the Siple Coast ice streams, the Evans Ice Stream; Figs. 1, 15a) others have tributaries that are relatively short and wide (e.g., Pine Island and Thwaites glaciers; Fig. 15b). In contrast, some modern ice streams do not form dendritic networks and, instead, only one trunk exists, commonly with diffuse lateral margins. Examples of these are the Sør Rondane and Belgica ice streams draining to the Princess Ragnhild Coast or the Ninnis Glacier in the George V Land (Figs. 1, 15c). Yet other ice streams do not feature one main trunk and, instead, display an anastomosing pattern of multiple fast-flow ‘channels’ (Fig. 15d). In some cases, especially with larger ice streams, combinations of the above types exist, with an intricate network of tributaries that display anastomosing patterns around isolated areas of slow-flowing ice and which feed a large broad trunk (Fig. 15e). In other cases, ice stream onset zones display convergence of flow towards a single downstream trunk that is often narrower and topographically defined (Fig. 15f). We also note that some downstream sections appear to show an indication of an inner and outer lateral margin, e.g. Thwaites Glacier (Fig. 15b).

***Fig. 15 here**, column width*

In comparison to modern ice streams (Fig. 15a-f), the shapes of LIS ice stream tracks can be divided into several similar classes, albeit with some notable exceptions:

(i) Dendritic ice streams where several tributaries, usually fed by several fjords, merge into a shelf-crossing trough (examples: the Nansen Sound Ice Stream, Smith Sound Ice Stream, Laurentian Channel Ice Stream; Figs. 3, 12, 15g);

(ii) Ice streams with the main trunk occupying a channel, with a convergent onset zone, possibly with few large tributaries (examples: the M’Clure Strait Ice Stream, Amundsen Gulf Ice Stream, Hudson Strait Ice Stream; Figs. 3, 15h, n);

(iii) Terrestrial ice streams with convergent onset zones and relatively narrow, winding trunk (ice streams on the southern Interior Plains or the Bear Lake Ice Stream; Figs. 9, 15i);

(iv) Ice streams whose whole track is represented by a convergent flow pattern – this type seems to consist entirely of deglacial ice streams (the Horton/Paulatuk Ice Stream, Haldane Ice Stream, Horn Ice Stream, Buffalo River Ice Stream, Bernier

Bay Ice Stream; Figs. 3, 9, 15k) albeit even some modern-day Greenland ice streams may attain this shape (Fig. 2);

(v) Hour-glass-shaped ice streams with no discrete tributaries and a convergent onset zone and divergent downstream end (the Dubawnt Ice Stream, Suggi Lake Ice Stream, James Bay Ice Stream, Kogaluk Ice Stream; Figs. 13, 15j);

(vi) Fan-shaped ice streams whose whole track is represented by a divergent fan-shape (the Hayes Lobe, Rainy Lobe, Red River Lobe; Figs. 13, 15l).

The groups of hour-glass-shaped (Fig. 15j) and fan-shaped ice streams (Fig. 15l) are absent among modern ice streams. However, ice streams operating in the Fennoscandian Ice Sheet during its retreat over southern Finland attained distinct fan shapes (Punkari, 1995; Boulton et al., 2001), albeit at a smaller scale, and, similarly to the Hayes and Rainy lobes, these were also deglacial ice streams terminating in shallow water. We discuss the longevity and significance of these types of ice streams in Section 5.5.

In terms of dimensions, ice streams in Antarctica and the LIS occur in a variety of sizes (Fig. 16). While small ice streams of only a few km in width and several tens of km in length occur in Antarctica and the LIS (feeding topographically defined outlet glaciers), the largest ice streams currently active in Antarctica are smaller than the largest Laurentide ice streams. While the longest of the Antarctic ice streams is the Recovery Glacier with ~900 km length (Fig. 1), the length of the largest LIS ice streams ranged between ~1300 and ~1000 km (Table 1, Fig. 2). However, it is important to note that the Antarctic Ice sheet was more extensive at the LGM when the overall lengths of Antarctic palaeo-ice streams were probably more comparable (e.g., ice streams in the Crary or Ronne troughs or at the Siple Coast may have reached 1300-1600 km at the LGM; Fig. 14; Livingstone et al., 2012). In addition, identifying the upstream limit is somewhat arbitrary for both modern and palaeo-ice streams. The upstream limit was defined as the uppermost spatially identifiable zone of enhanced velocity (i.e. bordered by slower moving ice) for Antarctic ice streams (in data from Rignot et al., 2011c); and for the measurement of LIS ice streams the decision was made on case-by-case basis and reflects the upstream limit of evidence for accelerating flow entering the ice stream system (Table 1). We also note that ice streams in smaller ice sheets tend to be shorter, but not proportional to the size ratio between the ice sheets: the largest ice stream in the Greenland Ice Sheet, the Northeast Greenland Ice Stream (Fig. 2), reaches ~700 km length, and the Baltic Sea Ice Stream of the Fennoscandian Ice Sheet could have reached a length of ~1000 km.

Table 1 here

Although the absolute size of the largest LGM ice streams of the LIS exceeds those operating at present in Antarctica, their length-to-width ratios are within the same range (Fig. 16). Interestingly, a distinct trend appears within the group of the LIS ice streams: deglacial ice streams have lower length-to-width ratios than ice streams draining ice to the LGM ice margin (Fig. 2, 16). This, together with anomalous shape of ice streams like the Hayes or Rainy lobes (Fig. 13), may indicate that the deglacial ice streams may have formed in reaction to dynamic or climatic forcing that does not occur at modern ice sheets.

Fig. 16 here*, column width*

In summary, it appears that the LGM velocity pattern of the LIS was organised in a similar way to the comparably sized modern Antarctic ice sheets. Under these conditions, most of the mass loss is delivered through ice streaming, rather than surface melt (Bamber et al., 2000; Shepherd et al., 2012). In contrast, the ice drainage pattern changed considerably during deglaciation of the LIS, when climatic conditions were likely to induce a greater proportion of surface melt (Carlson et al., 2008, 2009; Storrar et al., 2014). During deglaciation, the network of ice streaming was punctuated by shorter but broader ice streams that operated over the flatter interior regions and which have no modern analogues.

5.3. Marine versus terrestrial ice streams

For all present-day ice streams in Antarctica and Greenland, the large ice flux is calved directly into the ocean, and sometimes via large ice shelves (e.g. in Antarctica). However, the removal of ice from terrestrial ice stream termini is more enigmatic and, in most cases, these ice streams are associated with a lobate ice margin, which typically advances into lower elevation or warmer areas that help remove ice through ablation. Given that terrestrial ice streams did not perpetually advance, an obvious question is whether ablation rates at the downstream end are high enough to sustain continuous streaming flow or whether these ice streams represent a short-lived advance, followed by stagnation and ablation. These issues relate to other questions about the longevity and character of fast ice flow at terrestrial margins.

For a broad group of terrestrial ice lobes, the term surge has frequently been used (e.g., Clayton et al., 1985; Marshall et al., 1996; Kleman et al., 1997; Marshall and Clarke, 1997b; Evans and Rea, 1999; Evans et al., 1999; Kleman and Applegate, 2014). These fast flow features were expected to have undergone cycles of surging and quiescence, which has been supported by the reconstructed chronologies for the southern LIS margin indicating a fluctuating ice margin where individual ice lobes repeatedly advanced and retreated (Clark, 1994; Dyke et al., 2003; Mickelson and Colgan, 2003). They are also recorded by lateral moraines indicating low ice-surface slopes (Clayton et al., 1985), and by assemblages of landforms indicating stagnation of the surged lobes (Evans and Rea, 1999; Evans et al., 1999). Theoretical support for this mode of behaviour has come from the surging mechanism observed frequently at polythermal glaciers, where changes in the thermal regime at the bed and a build-up of subglacial water pressures lead to an abrupt onset of fast flow (Kamb et al., 1985; Kamb, 1987; Raymond, 1987).

In contrast, Patterson (1998) suggested that the lobes of the southern LIS margin operated not as short-lived surges, but as terrestrial ice streams that were sustained for longer time periods. She stressed the effects of the initial topography: ice would have preferentially been flowing in topographic lows where more subglacial meltwater was produced due to thicker ice, and fast ice flow would have further been induced by the fine sediments covering the floor of the shallow troughs. These initial conditions would have led to an establishment of a stable ice drainage network of the ice sheet comprising a number of persistent ice streams (Patterson, 1997; Patterson, 1998; Jennings, 2006).

To test whether terrestrial ice streams are able to persist, a simple calculation for mass flux can be done, and we use the dimensions of the James and Des Moines lobes at the southern margin. These were about 100 km wide at their downstream end and, because the ice thicknesses are not well constrained, two values, 500 and 1000 m, will be used. Assuming that the ice stream formed an ice lobe protruding from the adjacent non-streaming ice sheet margin with simplified dimensions of 300 long x 150 km wide, melt rates required to prevent the lobe advancing can be estimated from ice flow velocities within the ice stream. For a flow velocity of 1 km/year and an ice thickness of 500 m, the melt rate on the ice lobe would need to be about 1 m of ice per year (2 m for ice 1000 m thick). These values are well below the values modelled by Carlson et al. (2008, 2009) for the ablation area of the ice sheet during deglaciation. We therefore suggest that sustaining a terrestrial ice stream is less of a problem than might have been hitherto assumed and that the reconstructed short-lived surges (Evans and Rea, 1999, 2003) might have been characteristic mainly during the phase of ice retreat.

5.4. Controls on ice stream location

Where ice streams turn on and off in an ice sheet is an important control on the configuration and stability of ice sheets (Hughes, 1977; Stokes and Clark, 2001; Winsborrow et al., 2010). In this section, we discuss possible controls governing the location of ice streams within the LIS. In this regard, Winsborrow et al. (2010) identified several factors that may influence the location of ice streams: (i) topographic focusing, (ii) topographic steps, (iii) macro-scale bed roughness, (iv) calving margins, (v) subglacial geology, (vi) geothermal heat flux, and (vii) subglacial meltwater routing. In general, we find that almost all of the larger ice streams (with a notable exception of ice streams of the central Canadian Shield as well as central Alberta) exhibit at least partial topographic steering (Fig. 14) and that most of these ice streams also coincide with several other controls. This causes issues when trying to identify the primary control(s) on each individual ice stream, but we now discuss each of the potential controls and their likely importance across the population of ice streams in the LIS.

5.4.1. Topographic steering

Major topographic features exert a strong control on ice-flow pattern (e.g., Mathews, 1991). Fast ice flow in topographic troughs is supported by several processes (cf. Winsborrow et al., 2010): thicker ice reaches pressure melting point when surrounding ice on topographic highs is still frozen to the bed (Sugden, 1978; Hall and Glasser, 2003); thick ice under high pressure is more viscous than surrounding thinner ice (Clarke et al., 1977); and the floors of topographic lows are frequently covered by sediments that constitute a weaker bed than bedrock (e.g., Dowdeswell et al., 2004).

Topographic steering appears to be a dominant control on ice flow pattern both in the present-day ice streams of Antarctica and Greenland as well as in the LIS (cf. panels a and b in Fig. 14 and panels a-c in Fig. 7; Løken and Hodgson, 1971; Sugden, 1977, 1978; Denton and Hughes, 1981; England et al., 2006; Kessler et al., 2008). From the ice streams identified in the LIS, 55% were reconstructed based on the occurrence of glacial troughs and, of these, 89% display other evidence of their existence, such as a bedform imprint, IRD provenance, sedimentological evidence or the occurrence of sedimentary depo-centres (Fig. 6; Margold et al., in press). Whereas almost all of the ice streams draining the LIS during the LGM appear to be topographically controlled (Fig. 14), the degree of topographic control on ice stream location decreases during the deglaciation, and most of the larger deglacial ice streams show little relation to topography (Fig. 14). However, this is largely due to the fact that the ice sheet was retreating onto the central parts of the Canadian Shield, which is characterised by landscapes of low relief (Fig. 14). The exception is over the Interior Plains, where fast ice flow became increasingly steered by the topography during deglaciation (Figs. 9, 14; Ross et al., 2009; Ó Cofaigh et al., 2010).

5.4.2. Calving ice front

With the exception of ice streams at the southern margin, all LGM LIS ice stream systems were likely terminating in the ocean, despite lower sea levels (Fig. 2; England et al., 2006; Shaw et al., 2006; De Angelis and Kleman, 2007; Rashid and Piper, 2007; Todd et al., 2007; Li et al., 2011; Batchelor and Dowdeswell, 2014; Batchelor et al., 2013a; Batchelor et al., 2013b; Batchelor et al., 2014; Jakobsson et al., 2014). However, major uncertainties exist with regard to the existence and extent of ice shelves that could have exerted a buttressing effect and protected the ice stream termini from calving. There is also uncertainty regarding the extent of some ice streams on the continental shelf, such as the ice streams draining the Innuitian Ice Sheet to the NE (England et al., 2006, 2009). Although ice shelves might have prevailed in front of some marine-terminating ice streams, even during deglaciation (Hodgson, 1994; De Angelis, 2007; Stokes et al., 2009; Furze et al., 2013), ice calving is expected to have had an important role in the retreat of grounded ice from the channels of the CAA (De Angelis and Kleman, 2007; Stokes et al., 2009). A calving terminus, in combination with topographic steering and a weak bed, would have presented a strong stimulus for fast ice flow, as long as it was sustained by topography that permitted marine transgression and a propagation of the calving bay with the retreating ice front. It is also worth noting that calving is not restricted to marine margins. Proglacial lakes along the terrestrial margin may also have influenced the location of ice streams (e.g., Stokes and Clark, 2004). Thus, the water depth of these lakes was a critical parameter in controlling the occurrence of calving (Cutler et al., 2001) and it would be useful to determine the extent to which proglacial lakes accelerated deglaciation, e.g. using numerical modelling (Cutler et al., 2001).

5.4.3. Geology of the bed

Ice velocity is a function of stresses within the ice mass and the drag of the bed constitutes an important component of the force balance (Paterson, 1994). Thus, the geology of the bed in terms of the strength and roughness of the bedrock and the presence or absence of a layer of loose sediments can either facilitate or impede fast ice flow (Bell et al., 1998). Weak sedimentary rocks as well as thick sediment cover have been suggested to be conducive to fast ice flow and to exert a control on the occurrence of ice streams (Hicock and Dreimanis, 1992; Marshall et al., 1996; Anandakrishnan et al., 1998; Ó Cofaigh and Evans, 2001; Lian et al., 2003; Phillips et al., 2010). Indeed, regional geology appears to exerts a strong influence on the distribution of ice streams within the LIS (Fig. 8). The onset of the network of ice streams in the NW, W and SW sectors of the ice sheet (Fig. 9) is particularly striking, in that it occurs immediately down-ice from the abrupt transition between the Canadian Shield and the more deformable sedimentary substrates. Elsewhere, weaker beds composed of marine or lacustrine sediments have been suggested to facilitate ice streaming in the basins of the Great Lakes (Fisher et al., 1985; Hicock and Dreimanis, 1992), in Hudson Bay (Fisher et al., 1985; MacAyeal, 1993; Tarasov and Peltier, 2004), and in the channels of the CAA (Tarasov and Peltier, 2004). We also note, however, that whilst the Canadian Shield is likely to have offered a higher-friction substrate, and evidently appears to have supported fewer ice streams, it hosted several large, broad ice streams (see Section 5.2.) that were probably facilitated by basal sliding in association with elevated subglacial water pressures (Stokes and Clark, 2003a; Stokes and Clark, 2003b).

5.4.4. Meltwater at the bed

If subglacial water is present at sufficiently high pressures, it can greatly reduce effective pressures, which leads to a significant decrease in basal drag (e.g., Clayton et al., 1985; Kamb, 1987). In addition, if sediments present at the bed are saturated with water, they become more easily deformable (e.g., Blankenship et al., 1986; MacAyeal, 1989). Both of these processes have been confirmed by field-studies on present day Antarctic ice streams (Engelhardt et al., 1990; Engelhardt and Kamb, 1997; Engelhardt and Kamb, 1998; Kamb, 2001). Furthermore, spatial and temporal variations in the availability of subglacial meltwater are known to occur (Gray et al., 2005; Murray et al., 2008; Vaughan et al., 2008) and rerouting of meltwater has been suggested to cause speed-up, slowdown, or stagnation of ice streams in Antarctica (Alley et al., 1994; Anandakrishnan and Alley, 1997; Anandakrishnan et al., 2001; Wright et al., 2008; Beem et al., 2014). Surface melt-induced speed-up of ice streams in Greenland has also been hypothesised (Zwally et al., 2002; Parizek and Alley, 2004; Bartholomew et al., 2010), but the precise response of the subglacial drainage system is not always straightforward (Schoof, 2010; Sundal et al., 2011; Meierbachtol et al., 2013). Since large changes in the amount of supraglacially produced meltwater probably occurred on the deglaciating LIS (Carlson et al., 2008, 2009; Storrar et al., 2014), it can be assumed that similar changes affected the ice-flow pattern of the ice sheet and, potentially, the location of ice streams.

Increased availability of meltwater at the bed (either from subglacial or supraglacial sources), could thus have a significant influence on the location of fast ice flow and may help explain the large ice streams that operated in otherwise unfavourable conditions (e.g., with no topographic control and over a resistant bed) during deglaciation. Perhaps unsurprisingly, meltwater drainage pathways modelled by Livingstone et al. (2013) also correlate well with the majority of large topographic LIS ice streams (Fig. 17). We also note that the location of one of the few subglacial lakes hypothesised for the LIS (Great Slave Lake; Fig. 17; Christoffersen et al., 2008; Livingstone et al., 2013), lies immediately up-ice from an ice stream track (no. 175 on Fig. 9). This lake could thus have possibly promoted fast ice flow down-ice of its location in the manner suggested for Antarctic subglacial lakes (Siegert and Bamber, 2000; Bell et al., 2007).

***Fig. 17 here**, column width*

5.4.5. Macro-scale bed roughness, geothermal heat flux, and transverse topographic steps

In contrast to the controls on ice stream location discussed above, we observe relatively little evidence for the effects of geothermal heat flux, topographic steps transverse to the ice-flow direction, or bed roughness, which were also discussed by Winsborrow et al. (2010) as potential controls on ice stream location. Increased values of geothermal heat flux have been found to correlate with the onset zones of the Northeast Greenland Ice Stream (Fahnestock et al., 2001a, b) and the Siple Coast Ice Streams in Antarctica (Blankenship et al., 1993, 2001). Values of the geothermal heat flux show a large variation across the bed of the LIS (Fig. 18; Blackwell and Richards, 2004), in a pattern similar to the estimations for Antarctica, both in terms of spatial variations and absolute values (cf. Maule et al., 2005). Highest values, in excess of 100 mW/m^2, are reached in the southern Northwest Territories, and indeed, Brown (2012) suggested that, through its influence on subglacial melting, geothermal heat flux might have contributed to the development of ice streams in the NW sector of the LIS. However, these relationships are not straightforward. Elsewhere on the ice sheet bed, we note low geothermal heat flux values in the area of Hudson Bay and central Labrador, but whereas central Labrador exhibits correspondingly low ice streaming activity, several ice streams have been identified SW of Hudson Bay where the geothermal heat flux values are similarly low.

Fig. 18 here*, column width*

Macro-scale bed roughness (defined as ~1-100 km) has been shown to correlate with the ice velocity pattern of modern ice streams (e.g., Siegert et al., 2004; Rippin et al., 2011) and ice sheets (e.g., Bingham and Siegert, 2009). However, little systematic research to examine the influence of macro-scale bed roughness on the ice-flow pattern has been done for the Pleistocene ice sheets, which is perhaps surprising given the accessibility and data availability of palaeo-ice sheet beds. It has been observed that ice stream tracks in the SW sector of the LIS, outside of the Canadian Shield, are much smoother than the surrounding terrain (Evans et al., 2008, 2014). However, it is almost impossible to determine the cause and effect, and it is equally likely that the smooth bed results from ice stream flow, rather than caused it.

Similarly to bed roughness, topographic steps have received minimal attention in the case of the LIS and, where considered, they have been suggested to exert little influence on ice stream location (Brown, 2012). Indeed, the bed of the LIS had a much lower relief compared to Antarctica and Greenland (Fig. 14) and, consequently, transverse topographic steps were less likely to affect the character of ice drainage.

5.4.6. Summary

In summary, we find that topography appears to be the most influential control on the location of ice streams at the LGM (Fig. 14), with many topographic ice streams also terminating in the ocean and thereby possessing a calving margin. This is very similar to modern-day ice sheets in Greenland and Antarctica. During the first stages of deglaciation (18-11 cal ka BP), the southern and western margins of the ice sheet retreated over relatively deformable sedimentary substrates that appear to have facilitated a large number of sinuous ice streams that existed as dynamic networks (Fig. 8). The number of ice streams drops quite dramatically once the ice sheet retreated over much harder (and flatter) crystalline terrains of the Canadian Shield (Figs. 2, 8), suggesting that the underlying geology is also an important control. In this respect, our findings are in broad agreement with Winsborrow et al.'s (2010) hierarchy that suggests that topographic troughs, calving margins and soft beds are the most important controls on ice stream location. However, several ice streams 'turned on' during final deglaciation (10-8 cal ka BP), perhaps influenced by elevated subglacial water pressures, but with no obvious links to predicted meltwater drainage (Fig. 17) or physiographic controls. Some may have been influenced by calving into proglacial lakes, but we speculate that they were likely triggered by some form of mass balance (i.e. melt-induced) destabilisation linked to climate warming.

5.5. When did the ice streams operate?

Despite a comprehensive knowledge of the spatial extent of ice streams, our review indicates that there are few constraints on their temporal activity. This is a major gap in our understanding, because knowledge of when ice streams turned on and off is critical to an understanding of the response (and influence) of ice sheets to (on) the climate system. For example, to what extent was ice streaming driven by changes in ice sheet mass balance or localised physiographic controls (Section 5.4.)? Did ice streams turn on and off synchronously in response to, or during, major ocean-climate events (e.g., Heinrich events, meltwater pulses, abrupt warming or cooling)?

For some parts of the ice sheet, such as portions of the Keewatin and Foxe sectors, the timing of ice streaming has been broadly reconstructed using the most up to date ice margin chronology of Dyke et al. (2003; see Stokes and Clark, 2003b; Shaw et al., 2006; De Angelis, 2007; De Angelis and Kleman, 2007; Stokes et al., 2009; Brown, 2012). Preliminary data-model comparisons have also been used to inform our understanding of when some ice streams may have operated (Stokes and Tarasov, 2010; Stokes et al., 2012), but for most ice streams, there have been few attempts to constraint their activity using absolute dating methods (Winsborrow et al., 2004).

Ice streams tracks that extend to the maximum limit of the LGM ice sheet and/or extend across the continental shelf have generally been assumed to be active at the LGM (e.g., Kleman and Glasser, 2007) whereas those that lie well inside the LGM ice margin (e.g., the Dubawnt Lake Ice Stream) or those that deviate from the LGM ice-flow patterns (e.g., some of the smaller ice streams on Baffin and Prince of Wales islands; Figs. 2, 3) have generally been considered much younger (Stokes and Clark, 2003b; De Angelis, 2007; De Angelis and Kleman, 2007; Stokes et al., 2009). However, not all ice streams reaching the LGM ice margin necessarily operated simultaneously, which is highlighted by the varied timing of the maximum advance of the southern lobes (Clayton and Moran, 1982; Mickelson et al., 1983; Attig et al., 1985; Dyke and Prest, 1987a; Mickelson and Colgan, 2003; Kehew et al., 2005; Ross et al., 2009). The timing of operation is also uncertain for ice streams flowing across the continental shelf, in the Beaufort and Labrador seas and in Baffin Bay, beyond the maximum Late Wisconsinan limit of the ice sheet (Fig. 2). Indeed, the LGM ice margin has recently been re-drawn to the edge of the continental shelf in most areas (Shaw et al., 2006; Li et al., 2011; Lakeman and England, 2013; Jakobsson et al., 2014), but the timing of ice streaming remains uncertain, especially in terms of when they might have switched on in these settings: before, during or immediately after the LGM? Very little is known about pre-LGM ice streaming within the LIS. Both the landform record (Kleman et al., 2010) and terrestrial sediment dispersal (Shilts, 1980; Adshead, 1983; Prest et al., 2000) indicate that pre-LGM ice sheet geometry and ice flow patterns might have been distinctly different from the LGM and post-LGM periods, even though the results of low-resolution modelling studies show that some of the largest topographic ice streams may have operated during most of the ice sheet's existence (Stokes et al., 2012).

Important constraints on the timing of pre-LGM ice-streaming are likely to be recorded in ocean-floor sediments. In particular, major episodes of iceberg calving inferred from IRD records are able to span the entire late Pliocene and the Pleistocene (Bailey et al., 2010, 2012). A more detailed record is available for the late Pleistocene, and especially for the late Wisconsinan, and it shows a periodicity of IRD events likely related to LIS dynamics and with major ice fluxes operating with a roughly 7 kyr cycle (Heinrich events) that are in synchrony with the coldest peaks recorded in the Greenland ice cores (Hemming, 2004). The established timing of the Heinrich events is approximately 60, 45, 38, 31, 24, and 16.8 ka for H6 to H1, with a Heinrich-like event (H0) described during the Younger Dryas. The average duration of the Heinrich events is inferred to about 500 years (Andrews and MacLean, 2003; Hemming, 2004). The Heinrich oscillations overprint a finer pattern that shows an increase in the LIS dynamics that may reflect the cold peaks of the Dansgaard-Oeschger cycles (Bond and Lotti, 1995; Andrews and Barber, 2002). Given recent advances in sediment provenance techniques (Andrews and Eberl, 2012; Andrews et al., 2012), there would appear to be huge potential to make links between these IRD events and specific ice stream catchments.

Even though most research on IRD fluxes from the LIS has concentrated on the sediments deposited in the North Atlantic and traced back to the Hudson Bay and Strait region (Andrews, 1998; Hemming, 2004), major IRD events have also been linked to other ice streams, which released icebergs from the CAA to the Beaufort Sea and Baffin Bay, and from Labrador and Atlantic Canada to the Atlantic Ocean (Darby et al., 2002; Stokes et al., 2005; Rashid et al., 2012; Andrews et al., 2014; Simon et al., 2014). However, whereas some of this influx may be tentatively synchronous with Heinrich events (e.g., the activity of the M’Clure Strait Ice Stream: Darby et al., 2002; Stokes et al., 2005) other cyclic increases in ice stream activity do not correlate with this rhythm (Andrews et al., 2014). Furthermore, little connection has so far been established between the record of the LIS dynamics reconstructed from the ocean floor sediments (i.e., Heinrich events) and the terrestrial glacial landform and sedimentary record. To our knowledge the only exceptions are the advances of the Rainy Lobe in Minnesota, which Mooers and Lehr (1997) correlated with H2 and H; and the interpretation by Dyke et al. (2002) that changes in the ice sheet geometry over Labrador reconstructed by Veillette et al. (1999) might be linked to Heinrich event reorganisation.

Notwithstanding the lack of absolute age control, we can use the distribution, size and shape of ice streams to tentatively identify three different categories based on their temporal activity (see also Kleman et al., 2006; Kleman and Glasser, 2007). The first category, which we term '*persistent ice streams'*, are those reconstructed or assumed to have operated continuously, as long as their trajectories were preferential pathways for ice drainage, such as along major topographic troughs. Examples of these are the Amundsen Gulf Ice Stream (Stokes et al., 2009; Brown, 2012), ice streams draining Foxe ice across Baffin Island into Baffin Bay (Briner et al., 2006; De Angelis and Kleman, 2007; Briner et al., 2009), as well as possibly other marine-terminating ice streams draining the Innuitian Ice Sheet, the Labrador Ice Dome, and the ice complexes of Atlantic Canada (England et al., 2006; Shaw et al., 2006).

A second category, which we term '*recurrent ice streams'*, are those that have been interpreted to switch on and off in the same location. These would include the M'Clure Strait Ice Stream, which is thought to have been replaced by a short-lived ice divide, and then subsequently switched back on in the form of the smaller M'Clintock Channel Ice Stream (Hodgson, 1994; Clark and Stokes, 2001; Stokes et al., 2009), and possibly the James Lobe and the Des Moines Lobe, which have been reconstructed to advance and retreat several times during the Late Wisconsinan (Clayton and Moran, 1982; Dyke and Prest, 1987a). A long-frequency binge/purge oscillation throughout the glacial cycle, reflected in the Heinrich layers, has also been suggested for the Hudson Strait Ice Stream (Heinrich, 1988; Bond et al., 1992; MacAyeal, 1993; Alley and MacAyeal, 1994; Marshall and Clarke, 1997b; Calov et al., 2002; Robel et al., 2013).

A third category are those that only operated once and over a short time-scale (decades to a few centuries) and which we term '*ephemeral ice streams*' (after Kleman et al., 2006; including their category "transient rigid-bed ice streams"). These ice streams came into existence as a result of rapid changes in ice sheet geometry and transient conditions that promoted fast ice flow during deglaciation (Kleman et al., 2006; Stokes et al., 2009; Kleman and Applegate, 2014). Examples include the Dubawnt Lake Ice Stream or the Hayes Lobe (nos. 6 and 179 in Fig. 13) or small deglacial ice streams on Prince of Wales and Baffin islands (nos. 12, 101-103, 106-107, 118-120 in Fig. 3).

In summary, a number of large ice streams reached the LGM limit of the ice sheet and have thus been assumed to have operated during the LGM. However, our knowledge about the timing of ice stream operation within the LIS is uneven and incomplete. Whilst the temporal history of some ice streams is known in general outline, there have been few attempts to date ice stream activity in the LIS and this is a key area for future work to address. Nonetheless, the size and shape of ice streams suggests three main categories that we term persistent ice streams, recurrent ice streams, and ephemeral ice streams.

5.6. Stability of ice drainage network

In relation to the previous section, it is important to consider the temporal and spatial stability of the ice stream drainage network, which we broadly define as the pattern and spacing of ice streams. Research on contemporary ice sheets is heavily focussed on measuring and modelling changes in ice stream velocity, thinning and terminus positions (Joughin, 2002; Joughin et al., 2004, 2008; Nick et al., 2009; Miles et al., 2013; Nick et al., 2013) and yet we have little context for understanding what changes might take place over much longer centennial to millennial time-scales, e.g., will ice streams persist or will other ice streams switch on or off? Knowledge of palaeo-ice streams, however, should allow us to answer some of these questions and assess how stable the ice stream drainage network might be within a deglaciating ice sheet.

In the LIS, an obvious control on the ice stream network is topography (see Section 5.4.). High relief coasts overrun by the ice sheet (such as NW Ellesmere Island and NE Baffin Island) exhibit a regular pattern of ice drainage organisation where several fjords feed into a shelf-crossing trough. This organisation with regular spacing between the cross-shelf troughs and the highly over-deepened trough heads requires a prolonged time for formation (Kessler et al., 2008), which attests to a relatively stable ice drainage network in these portions of the ice sheet, probably over several glacial cycles. Analogous settings existed in the Pleistocene Cordilleran and Fennoscandian ice sheets and in the Greenland and Antarctic ice sheets (Fig. 7). It is interesting to note, however, that along these heavily incised coasts, there appears to be a clear preference/organisation of ice stream spacing, which presumably reflects the interaction of the catchment areas that feed individual fjords. For example, Fig. 7a shows numerous (8-9) relatively short and closely spaced cross-shelf troughs emanating from the coast of Baffin Island, whereas across Baffin Bay, the ice streams from west Greenland carved much larger troughs that were spaced further apart. This organised pattern and spacing has rarely been scrutinised in contemporary or palaeo-ice sheets, but hints at a regulatory role of ice streams in these regions where the potential for additional ice streams to switch on and off is, presumably, limited. Of course, this does not preclude temporal variations in ice flux from individual ice streams, perhaps through short-term bathymetric controls or changes in the size or slope of the ice stream catchments (Briner et al., 2009; Jamieson et al., 2012; Joughin et al., 2014; Rignot et al., 2014; Stokes et al., 2014).

Elsewhere in the ice sheet, there is evidence that the drainage network of ice streams was far more dynamic, typically in lower relief areas, such as across the Canadian Prairies (e.g., Evans et al., 2008; Ross et al., 2009; Ó Cofaigh et al., 2010; Evans et al., 2014), and in Labrador/Ungava (Kaufman et al., 1993; Clark et al., 2000; Jansson et al., 2003). The interaction between neighbouring ice streams has been observed during deglaciation (Ross et al., 2009; Ó Cofaigh et al., 2010; Evans et al., 2014). Even in some of the moderately high relief settings, changes in ice catchments might drive changes in ice stream activity. For example, at the NW margin of the ice sheet, Stokes et al. (2009) noted periods when neighbouring ice streams appeared to behave in synchrony, such as during the retreat of the M’Clure Strait and the Amundsen Gulf ice streams between 15.2 and 14.1 cal ka BP, for which a dominance of external forcing was inferred. However, they also identified times when the ice streams behaved differently, and thought to reflect internal dynamics of the ice stream catchments (Stokes et al., 2009). De Angelis (2007) also stressed the importance of nonlinear processes involved in internal ice sheet dynamics, where a large reaction may be triggered by a minor change in external conditions or ice sheet configuration. An illustration of this might be an inference of Stokes et al. (2009) that the quiescence of the M’Clure Strait Ice Stream was caused by its previous rapid retreat into deeper and wider Viscount Melville Sound, which led to a thinning and a subsequent freeze-on of the ice mass (cf. Christoffersen and Tulaczyk, 2003; Beem et al., 2014), and to profound changes in the configuration of the ice sheet sector.

Thus, although there has been limited work on the stability of ice stream drainage networks at millennial time-scales, our synthesis from the LIS appears to show stable and regularly spaced networks in areas of high relief, but with the potential for much more dynamic changes to occur over low relief areas (Jansson et al., 2003; Ross et al., 2009; Ó Cofaigh et al., 2010). This “switching” behaviour is likely driven by a number of factors (see also Winsborrow et al., 2012), including changes in topography and geology as the ice sheet retreats (Dowdeswell et al., 2006; Stokes et al., 2009), competition and interaction between neighbouring catchments (in terms of both ice and subglacial meltwater; Payne and Dongelmans, 1997; Anandakrishnan et al., 2001; Conway et al., 2002; Greenwood and Clark, 2009) and, potentially, external climate triggers (De Angelis and Kleman, 2007; Stokes et al., 2009).

5.7. What role did ice streams play in ice sheet mass balance during deglaciation?

A further interesting question that relates to the stability of the ice stream drainage network relates to the role of ice streams during ice sheet deglaciation. In contemporary ice sheets, ice streams account for between 50% (Greenland) and up to 90% (Antarctica) of the ‘dynamic’ mass loss, with the remaining accounted for melting (supraglacial or basal, e.g., under ice shelves; Bamber et al., 2000; van den Broeke et al., 2009). To date, however, there have been no empirical estimates for the potential flux from ice streams in the LIS, or any other palaeo-ice sheet, either for the LGM or for different stages of deglaciation. Did the percentage of dynamic mass loss remain constant throughout the deglaciation or did it increase or decrease? We illustrate these three simple scenarios in Fig. 19 and suggest that determining their likelihood is likely to represent a significant advance in understanding ice sheet response to major changes in climate. If we knew the answer, it might tell us whether ice stream activity across an ice sheet is predictably and closely related to external climate forcing, or whether it might accelerate deglaciation far quicker than might be expected from climate forcing alone. The latter would have major implications for our predictions of modern-day ice sheets and the time-scales and magnitude of future sea level rise.

***Fig. 19 here**, column width*

Although the interplay between the effects of external forcing and internal dynamics during the LIS deglaciation was undoubtedly highly complex, there are some hints of internally driven instabilities that might be unrelated to climate forcing. The binge/purge explanation for Heinrich events (MacAyeal, 1993), if correct (see discussion in Hemming, 2004), would suggest that strong ice-dynamical mechanisms operated at least in some sectors of the ice sheet. Furthermore, a number of ice stream tracks fit the definition of ice stream "singular events" (see Kleman and Applegate, 2014), and might have thus been responsible for a substantial draw-down of the LIS ice mass in their respective sectors. Much like surge-type glaciers, these ice streams may have been influenced by long-term climate warming, but the precise timing of the response may have been more closely linked to changes in the distribution and pressure of subglacial melwater. Ice streams that we suggest might fall into this category include the Dubawnt Lake Ice Stream, the Hayes and Rainy lobes, some of the Ungava fans, the James Bay Ice Stream, the Maguse Lake Ice Stream (Fig. 13), and a number of smaller ice streams, particularly in the CAA. Indeed, during final deglaciation, numerical modelling studies (Beget, 1987; Carlson et al., 2008; 2009) as well as analyses of the landform record (Storrar et al., 2014) appear to indicate intense surface ablation.

To provide a first order estimate on the potential role of ice streams during the LGM and later during deglaciation, we calculate the percentage of the margin intersected/drained by ice streams at three time steps and compare it against the present-day Antarctic ice sheets. We estimate that the Antarctic ice sheet margin that is streaming (using the definition from Section 3.) is around 30% of its circumference (Fig. 20 a). The figure for the LGM LIS is 32% (Fig. 20 b). Despite reaching similar numbers for the present-day Antarctica and the LGM LIS, we note an important difference: the result for the LIS is derived from a much smaller number of relatively large ice streams compared to Antarctica. Furthermore, it is likely that in some areas of the ice sheet, we overestimate ice streaming activity at the LGM by adopting a simplified approach to the timing of ice stream operation, e.g., the ice streams at the southern margin may not have operated simultaneously (Kehew et al., 2005; Ross et al., 2009). On the other hand, it is likely that we are missing small ice streams of the size that we can still clearly distinguish in the Antarctic ice velocity data (Figs. 1, 20). Interestingly, for two subsequent time steps (~12 cal ka and ~10 cal ka) we estimate significantly lower percentages of the ice margin to be streaming: 15% and 12%, respectively (Fig. 20 b). This may reflect the fact that some of the potential controls/triggers for ice streaming were lost when the ice sheet retreated onto a hard bed (Clark, 1994; Marshall et al., 1996; Stokes et al., 2012), for example, soft sediments and a calving margin (see Section 5.4.).

It is also likely that numerical modelling could shed some light on this important issue. In a preliminary assessment of ice stream activity during LIS build-up from ~120 ka, Stokes et al. (2012) found a strong correlation between the size of the ice sheet and the relative role of dynamic mass loss (ice stream activity), an observation that is in agreement with the sedimentological record of ice dynamics on the ocean floor (Kirby and Andrews, 1999; Hemming, 2004). However, that model is limited by the relatively coarse grid size (that is unable to resolve narrow ice streams) and the use of the shallow-ice approximation, which may be unable to resolve the dynamics of ice streaming. To make further progress on this important issue therefore, requires (i) the deployment of numerical models with better grid resolution and higher order physics, and (ii) a concerted effort to constrain the temporal activity of ice streams through time (see Section 5.5.).

Fig. 20 here*, column width*

5.8. Future work

A comprehensive inventory of ice streams in the LIS is a powerful tool for improving our understanding of the controls on ice stream activity and their role in ice sheet mass balance and stability. Based on our synthesis and discussion in Sections 5.1.-5.7., we briefly highlight some key areas that future work might address:

- *Improved dating of ice stream operation* (see Section 5.5.). Constraining the timing of individual ice streams is a key requirement for answering important questions related to their activity and role in ice sheet mass balance, e.g., was ice stream activity linked to major climate events or transitions, or did they play a more regulatory role? Is there evidence for near-synchronous activation or deactivation of ice streams? Previous work has tended to use existing pan-ice sheet margin chronologies (Dyke et al., 2003) for specific regions (see De Angelis, 2007; De Angelis and Kleman, 2007; Stokes et al., 2009), but this has never been applied across the whole ice sheet. Moreover, there is a clear need for concerted efforts to specifically date palaeo-ice stream tracks, especially in the western and northern sectors of the ice sheet.
- *Provenance studies of IRD records from ocean-floor sediments*. In relation to the previous point, the timing of several marine-terminating ice streams along the

northern and eastern margin of the ice sheet might be further constrained by IRD records in the North Atlantic and Arctic Oceans (e.g., Rashid et al., 2012). These records have the added advantage of being able to extend our knowledge of their activity prior to the LGM, where terrestrial evidence is scarce (Stokes et al., 2012).

– *Criteria for examining hard-bedded ice streams* (see Section 5.1.). Our knowledge of ice stream geomorphology is mostly gleaned from those that operated over soft, unconsolidated sediments, where the bedform imprint is most obvious, e.g., mega-scale glacial lineations (Fig. 5). There is much more uncertainty about the geomorphological imprint of ice streaming over hard beds, although some putative criteria are emerging (Bradwell, 2005; Bradwell et al., 2008; Roberts and Long, 2005; Roberts et al., 2010; Eyles, 2012; Eyles and Putkinen, 2014). Further work could usefully focus on differentiating the imprint of slow versus fast flow over hard bedrock surfaces, further informed by geophysical surveying of active ice streams in these settings (see Bingham et al., 2010; Jezek et al., 2011; Jezek et al., 2013; Morlighem et al., 2013). Once identified, the geomorphology of hard-bedded ice streams might also allow inferences to be made about the flow mechanisms of these ice streams and the efficacy of glacial erosion in these settings, which affects bed roughness. Indeed, there is huge potential to use palaeo-ice stream settings, on both hard and soft beds, to examine the influence of bed roughness on ice sheet flow patterns, something that is being investigated on contemporary ice sheets/streams, despite the difficulty of obtaining high resolution data (Bingham and Siegert, 2009; Rippin et al., 2011, 2014). Measurements of bed roughness on palaeo-ice stream beds might be a powerful interpretative tool for these modern-day ice stream studies (Gudlaugsson et al., 2013).

- *Estimates of ice fluxes from palaeo-ice streaming*. In order to examine the role of ice streams in palaeo-ice sheet mass balance and stability (see Section 5.7.), it is necessary to estimate the potential magnitude of their ice flux through time. This requires better dating of palaeo-ice streams (see above), but also an improved understanding of their ice thickness and velocity, which would allow estimates of their ice flux. Due to the large uncertainties, these issues are often neglected in palaeo-ice stream studies, but future work could investigate techniques to better constrain ice thicknesses and velocities, perhaps using modern analogues and/or numerical modelling (Golledge et al., 2008; Stokes and Tarasov, 2010; Golledge et al., 2012).
- *Successful replication of palaeo-ice streaming in numerical ice sheet models*. Future predictions of contemporary ice sheet dynamics are heavily reliant on numerical ice sheet models. Our confidence in their ability to predict the behaviour of ice streams will gain confidence from their ability to replicate observations of past ice stream behaviour. Much progress has been made in attempting to model the behaviour of individual ice streams in both palaeo and modern settings (Boulton et al., 2003; Boulton and Hagdorn, 2006; Jamieson et al., 2012; Nick et al., 2013; Lea et al., 2014), but there have been very few attempts to compare model output against ice stream locations at the ice sheet scale. Stokes and Tarasov (2010) did this for the LIS, based on a much smaller inventory of ice streams, and found that most major topographic ice streams were captured, but that the model was not always able to resolve terrestrial ice streams. This is likely to reflect the inability of that model to fully capture the role of subglacial hydrology in generating fast flow over relatively flat beds, and this is a key area for future work to address.

6. Conclusions

This paper presents a comprehensive review and synthesis of ice streams in the Laurentide Ice Sheet, based on a new mapping inventory that includes previously hypothesised ice streams and includes a concerted effort to search for others from across the entire ice sheet bed (Margold et al., in press). The inventory includes 117 ice streams and, despite some subjectivity in identifying them over hard bedrock areas, it is unlikely that any major ice streams have been missed. At the LGM, Laurentide ice streams formed an ice drainage pattern that bears close resemblance to the present day velocity patterns of the similarly-sized Antarctic Ice Sheet (including both the East and West Antarctic Ice Sheets). Large ice streams had extensive onset zones and were fed by multiple tributaries. There is also similarity between the Laurentide and Antarctic/Greenland ice sheets when ice drained from or through regions of high relief onto the continental shelf, and where ice streams show a degree of spatial self-organisation which has hitherto not been recognised. However, the size of the largest Laurentide ice streams surpassed the size of ice streams currently operating in Antarctica.

Similar to modern ice sheets, most large ice streams in the LIS appear to have been controlled by topography, but there are zones along the western and southern margin where ice streams were spatially more dynamic and existed in sinuous tracks and show clear switches in trajectory during deglaciation. More generally, we note that the underlying geology exerts an important control on the pattern and density of ice streams, as noted in previous work (Fisher et al., 1985; Marshall et al., 1996; Clark, 1994). As the ice sheet retreated onto its low relief interior, several ice streams operated that show no correspondence with topography or underlying geology. Their location may have arisen from localised build-up of pressurised subglacial meltwater, and they differed from most other ice stream tracks in having much lower length-to-width ratios, often displaying convergent ice-flow pattern along their whole trajectory. Perhaps because all modern ice streams are marine-terminating, the feasibility of sustaining ice streams with a land-terminating margin is questionable, but we suggest that realistic melt rates of 1-2 m of ice per year are sufficient to ablate ice from a large, thin, divergent lobe that is fed by persistent rapid ice flow.

The timing of a handful of ice streams has been investigated through a proxy record of IRD sediments on the ocean floor (e.g., Heinrich events), which hints that the activity of some ice streams is linked to abrupt climate changes recorded in the Greenland ice cores (Bond and Lotti, 1995; Darby et al., 2002; Andrews and MacLean, 2003; Stokes et al., 2005). However, there is minimal dating control for the vast majority of ice streams in the LIS. Time-dependent ice sheet reconstructions that incorporate ice stream activity have only been carried out for some sectors of the ice sheet, such as the CAA (De Angelis and Kleman, 2005, 2007; Stokes et al., 2009), Atlantic Canada (Shaw et al., 2006), and the Great Lakes region (Kehew et al., 2005, 2012), whereas for other regions the timing of ice streams has rarely been investigated (e.g., the Interior Plains).

In terms of the stability of the ice stream drainage network, high relief areas fixed ice streams in topographic troughs, but it is clear that other ice streams switched on and off during deglaciation, rather than maintaining the same trajectory as the ice margin retreated. We note evidence for dynamic adjustments and reactions of the ice drainage network to changes in ice geometry and external forcing during the deglaciation. These include some of the late glacial ice streams, which appear to be local instabilities during an otherwise predictable ice margin recession, and with the potential of substantial draw-down of ice in the respective regions (Kleman and Applegate, 2014). This type of ice stream has no modern analogue, but is likely to occur if and when modern-ice sheet margins are forced to retreat onto flat interior regions in a warming climate (e.g., parts of Greenland). The extent to which changes in the ice stream drainage network represent a simple readjustment to a changing mass balance driven by climate, or internal ice dynamical feedbacks unrelated to climate (or both) is largely unknown and represents a key area for future work to address.

We provide a first order estimate of the changes in ice stream activity during deglaciation. The percentage of ice margin that was streaming at the LGM is remarkably similar to that for the modern Antarctic ice sheets (~30%), whereas this percentage drops significantly during the LIS deglaciation (to 15% at ~12 ka and just 12% at ~10 ka). This is consistent with recent modelling studies (e.g., Carlson et al., 2008, 2009) that have suggested an increasing role of surface melt during deglaciation, although those studies did not investigate the potential for 'dynamic' losses. This is a key area for future work to address and we suggest that dating of ice streams is an urgent priority. Such dating would help answer some key questions relating to the role of ice streams in ice sheet mass balance and whether they have potential to accelerate deglaciation beyond that which might be expected from climate forcing alone.

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Tables:

- Table 1. The longest Laurentide ice streams

Supplementary data:

- Information and literature for LIS ice streams

Figures:

Fig. 1. Ice flow of the Antarctic ice sheets from Rignot et al. (2011c). Ice sheet grounding line (from Rignot et al., 2011a) is shown in orange (medium grey if viewing a black and white version of the manuscript). Ice streams mentioned in the text are marked, as well as the location of Fig. 7b and c. This figure is drawn to the same scale as Fig. 2.

Fig. 2. Ice streams of the Laurentide Ice Sheet (LIS) drawn after Margold et al. (in press). LIS extent is shown for the Last Glacial Maximum (LGM) and at 10.2 cal ka BP, from Dyke et al. (2003). Note that the LIS has recently been shown to extend to the continental shelf at the LGM in many regions (e.g., Briner et al., 2006; Shaw et al., 2006; Kleman et al., 2010; Lakeman and England, 2012, 2013; Jakobsson et al., 2014; Nixon and England, 2014). The locations of Figs. 3, 9, 12, 11, and 13 are marked by black rectangles. Present-day glaciation is in light turquoise with ice margins in thin purple line (light grey with a dark grey margin if viewing a black and white version of the manuscript). Ice flow velocity for the Greenland Ice

Sheet is reproduced from the data released by Joughin et al. (2010a); data coverage is not complete; missing data is shown in white. Present-day coastline and administrative boundaries are drawn in grey (applies also for subsequent figures). This figure is drawn to the same scale as Fig. 1.

Fig. 3. Palaeo-ice streams in the Canadian Arctic Archipelago (see Fig. 2 for location). Ice flow pattern of this ice sheet sector is described in Section 4.1. and more information and evidence for individual ice streams is available in Supplementary data. Abbreviations: BP – Boothia Peninsula, CB – Committee Bay, CI – Coats Island, DS – Dease Strait, PWI – Prince of Wales Island, RI – Rae Isthmus, RGSI – Royal Geographical Society Islands, SI – Somerset Island (see Table 1). Location of panel (a) in Fig. 7 is marked by a black rectangle. Boundary of the Canadian Shield is marked by a pink stippled line (medium grey if viewing a black and white version of the manuscript). LIS extent is shown for the Last Glacial Maximum (LGM) and at 10.2 cal ka BP after Dyke et al. (2003), but note that it has recently been shown to extend to the continental shelf in many regions (e.g., Kleman et al., 2010; Lakeman and England, 2012, 2013; Jakobsson et al., 2014). Ice streams of a neighbouring province (with respect to our geographical subsections of Section 4.) are in grey and are found in separate figures.

Fig. 4. Panchromatic Landsat image of southern Prince of Wales Island (reprinted with permission from De Angelis, 2007). Elongated bedforms depict changing ice flow directions. Boundaries of fast ice flow are indicated by a shear margin moraine (see Dyke and Morris, 1988; Stokes and Clark, 2002) running S-N across the centre of the image and by the outline of a sediment dispersal train in the case of the Transition Bay Ice Stream that flowed in easterly direction in the lower right part of the image.

Fig. 5. Highly elongated mega-scale glacial lineations (MSGL) on the bed of the Dubawnt Lake Ice Stream. (A) Landsat imagery (path 039, row 015) of a portion of a central trunk of the ice stream, (B and C) oblique aerial photographs of parts of the image in panel A; photographs: C. R. Stokes, panels A-C reprinted from Stokes et al. (2013) with authors' permission. (D) MSGL identified on the bed of the Rutford Ice Stream in Antarctica (see Fig. 1 for location) compared to MSGL on the bed of the Dubawnt Lake Ice Stream (E); panels D-E reprinted from King et al. (2009) with authors' permission.

Fig 6. Types of evidence available for individual ice streams: bedform imprint (full bedform imprint in dark blue, discontinuous and isolated bedform imprint in lighter shades of blue [shades of grey if viewing a black and white version of the manuscript]); broad-scale topography (glacial troughs); sedimentary depo-centre at the edge of the continental shelf; ice-rafted debris (IRD); and sediments conducive to fast ice flow. More information and evidence for individual ice streams is described in Supplementary data.

Fig. 7. High-relief coasts modified by selective linear erosion under ice sheet glaciation. (a) Topographic map of Foxe Basin, Baffin Island and Baffin Bay (IBCAO data from Jakobsson et al., 2000; location marked in Fig. 3). Glacial troughs on the continental shelf NE of the Baffin Island coast were incised by ice draining from the Foxe Dome across Baffin Island; ice-flow directions are shown by black arrows. Note the difference in the size and spacing of glacial troughs and sediment bulges along the Baffin and Greenland sides of Baffin Bay, which hints at some form of spatial self-organisation, i.e. many narrow and closely-spaced ice streams versus fewer, broader ice streams spaced further apart. Present-day glaciation is marked in a semi-transparent blue (grey if viewing a black and white version

of the manuscript). (b) Subglacial topography of Dronning Maud Land and the Princess Astrid Coast in East Antarctica. Bedmap2 data from Fretwell et al. (2013) are significantly less detailed than the data used in panel a; location is marked in Fig. 1. (c) Ice-flow pattern for the area shown in panel b (data from Rignot et al., 2011c) – note the topographic steering of the major ice streams. All panels are drawn to the same scale and with the same hypsometric colour scale.

Fig. 8. Rock types on the Laurentide Ice Sheet bed. Ice streams are drawn by arrows; those inferred to be active at the Last Glacial Maximum are in pink; deglacial ice streams or those with unknown age are in purple (lighter and darker grey, respectively, if viewing a black and white version of the manuscript). Note the increased occurrence of ice streams beyond the edge of the Canadian Shield (regions in north-central Canada built of crystalline rocks).

Fig. 9. Ice streams in the region of the Interior Plains (see Fig. 2 for location). Ice flow pattern of this ice sheet sector is described in Section 4.2. and more information about individual ice streams is available in Supplementary data. Boundary of the Canadian Shield is marked by a pink stippled line (medium grey if viewing a black and white version of the manuscript). Abbreviations: CH – Cameron Hills, CM – Caribou Mountains, BM – Birch Mountains.

Fig. 10. Evidence for fast ice flow in the region of northern Interior Plains. (a) Broad troughs seen in a DEM-derived image draped with Landsat Image Mosaic of Canada. The trough floors are largely devoid of a continuous pattern of glacial lineations. However, isolated patches of extremely well-developed mega-scale glacial lineations occur both on the trough floors and on the slopes and upper surfaces of the intervening plateaux. Although the glacial

troughs define an ice stream configuration in the area (panel b – cf. Fig. 9 for location), streamlined terrain on the plateau surfaces (classified as ice stream fragments; Margold et al., in press) indicates a stage of fast ice flow that was not controlled by topography. (c) Streamlined surface of the Cameron Hills seen in a false-colour composition of SPOT satellite images (see panel a for location; scenes used: S4_11650_6004_20090901, S4_11709_5937_20090605, S4_11750_6004_20060623). Note the contrast between the slopes of the trough that display indistinct lineations along the direction of the trough and a heavily streamlined surface of the plateau, with the direction of streamlining independent of the trough orientation (see panel d for a close-up of the plateau surface edge).

Fig 11. Ice streams in the region of the Great Lakes (see Fig. 2 for location). Ice flow pattern of this ice sheet sector is described in Section 4.3. and more information about individual ice streams is available in Supplementary data. Boundary of the Canadian Shield is marked by a pink stippled line (medium grey if viewing a black and white version of the manuscript).

Fig. 12. Ice streams in the region of the Atlantic seaboard (see Fig. 2 for location). Ice flow pattern of this ice sheet sector is described in Section 4.4. and more information about individual ice streams is available in Supplementary data. Boundary of the Canadian Shield is marked by a pink stippled line (medium grey if viewing a black and white version of the manuscript). LIS extent is shown for the Last Glacial Maximum (LGM) and for 10.2 cal ka BP after Dyke et al. (2003), but note that a greater LGM extent has recently been inferred for the continental shelf (Shaw et al., 2006). PDM – Pointe-des-Monts (see Table 1).

Fig. 13. Ice streams in the region of the Canadian Shield (see Fig. 2 for location). Ice flow pattern of this ice sheet sector is described in Section 4.5. and more information about

individual ice streams is available in Supplementary data. Boundary of the Canadian Shield is marked by a pink stippled line (medium grey if viewing a black and white version of the manuscript).

Fig. 14. Topography of ice sheet beds. (a) Present-day (isostatically uplifted) topography of the area covered by the Laurentide Ice Sheet. Simplified Last Glacial Maximum extent is drawn after Shaw et al. (2006), Kleman et al. (2010), Jakobsson et al. (2014). Ice streams are drawn by arrows; those inferred to be active at the Last Glacial Maximum are in pink; deglacial ice streams or those with unknown age are in purple (lighter and darker grey, respectively, if viewing a black and white version of the manuscript). Present day glaciation is marked in white. (b) Subglacial topography of Antarctica (data from Fretwell et al., 2013); ice streams are drawn in pink, ice shelves in transparent blue (shades of grey if viewing a black and white version of the manuscript). Troughs on the continental shelf mentioned in the text are marked. Both panels are drawn to the same scale and use the same hypsometric colour scale.

Fig. 15. Shapes of ice streams. Shapes of present-day Antarctic and Greenlandic ice streams (all data from Rignot et al., 2011c; panel n reproduced from data by Joughin et al., 2010a) compared to Laurentide palaeo-ice streams (Margold et al., in press). Different ice stream shapes discussed in the text are drawn, all to the same scale. See Figs. 1, 2, 3, 9, 13 for locations. Similarities occur between modern ice streams in Antarctica and ice streams of the Laurentide Ice Sheet: branching and anastomosing patterns are present in both groups. Hour-glass-shaped and fan-shaped ice streams (panels j and l) are absent among modern ice streams, but occurred in the Fennoscandian Ice Sheet during deglaciation (redrawn from Boulton et al., 2001).

Fig. 16. Length-to-width ratio of large ice streams in present-day Antarctic and Greenland ice sheets and in the Laurentide Ice Sheet. *The Thwaites Glacier is shown twice because it appears to have an inner and outer lateral margin on both sides (see text in Section 5.2.).

Fig. 17. A composite of meltwater drainage pathways (blue to purple thin lines [medium grey if viewing a black and white version of the manuscript]) derived from the calculation of the hydraulic potential surface at the ice-sheet bed for 293 modelled ice-surface geometries during the period of 32-6 ka BP (Livingstone et al., 2013; reproduced with permission) alongside reconstructed ice streams (Margold et al., in press) drawn in orange/darker grey (LGM) and yellow/lighter grey (deglacial). Identified subglacial lake evidence in Christie Bay, Great Slave Lake (Christoffersen et al., 2008), is marked by a black star. Simplified LGM ice sheet margin is in pink (medium grey if viewing a black and white version of the manuscript). Note the correspondence between the modelled drainage locations and reconstructed ice stream tracks at times of maximum ice extent. However, ice streaming conditioned by the presence of meltwater cannot be directly inferred from this because topography plays a large role both in the modelled meltwater drainage pathways and in the location of ice streams.

Fig. 18. Present-day geothermal heat flux for the area formerly covered by the Laurentide Ice Sheet (modified from Blackwell and Richards, 2004). Simplified Last Glacial Maximum extent is drawn after Shaw et al. (2006), Kleman et al. (2010), Jakobsson et al. (2014). Ice streams are drawn by arrows; those inferred to be active at the Last Glacial Maximum are in pink; deglacial ice streams or those with unknown age are in purple (lighter and darker grey, respectively, if viewing a black and white version of the manuscript).

Fig. 19. Conceptual scenarios for the percentage of dynamic mass loss in the Laurentide Ice Sheet during deglaciation. Three possible scenarios are drawn: (i) the percentage of mass loss delivered by ice streams remained stable; (ii) the percentage of mass loss delivered by ice streams decreased during deglaciation, with a proportionally increasing contribution from surface melt, (iii) the percentage of mass loss through ice streams increased during deglaciation, perhaps hinting at non-linear feedbacks accelerating mass loss beyond that which might be expected from climate forcing alone.

Fig. 20. Percentage of the streaming margin calculated for Antarctica (a) and the Laurentide Ice Sheet (b), using the definition of an ice stream as spatial partitioning of ice flow (see Section 3.). Streaming margin was mapped manually for Antarctica from data by Rignot et al. (2011, c) and ice streams reconstructed by Margold et al. (in press) were used for the Laurentide Ice Sheet. The Laurentide ice margin is straightened for the inclusion of the large lobes formed by terrestrially terminating ice streams. Note that the coarseness of the method used implies that the results can be used as a first estimate only.